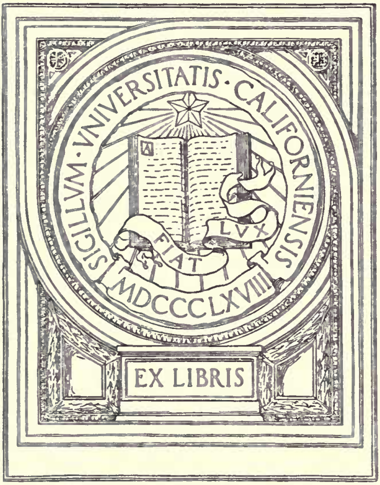
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MINERAL DEPOSITS

BY

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PREFACE TO THE SECOND EDITION

The last few years of scientific progress frequently loom up in magnified proportions. This is a view that may be corrected by a proper perspective but to the author of a text-book anxious to carry his readers up to date it is certainly disconcerting to measure the rapid accumulation of new facts and theories.

In spite of the disturbed condition of the world during the last six years, students of minerals and mineral deposits have made many important contributions to science. Among these may be counted the investigations bearing on magmatic and contact-metamorphic ore deposits, on problems of oxidation and supergene sulphides, and particularly the application of metallographic methods to ores by which the complexity of metallic replacements has been revealed.

In this second edition all of the chapters have been revised and those on deposition of minerals, contact-metamorphism, oxidation and sulphide enrichment, have been largely rewritten. A discussion of metallogenetic epochs has been added, as well as an index by elements which will enable the student to coördinate rapidly the deposits of any given metal. Many new illustrations have been introduced. Owing to war conditions it has not always proved easy to obtain late statistical data. In case of some European countries the latest figures available date from 1913.

The progress of science emphasizes the difficult task of condensation. In the present volume some less essential descriptions have been omitted so that its bulk has not been greatly increased. No one knows better than the author that errors may easily creep in and corrections and suggestions will be gratefully received.

The genetic arrangement has been preserved throughout. While it is realized that this may make the study less easy for beginners, it is believed that any merit that the book may have is due to this mode of treatment.

Cambridge,
April, 1919.

PREFACE TO FIRST EDITION

Mineral deposits are usually classified and described by the metals or the substances which they contain; for instance, deposits of copper are described together, with little or no effort to separate them into genetic groups. Where a genetic treatment has been attempted it appears to me to have failed in not giving due weight to the physical conditions attending the genesis. Furthermore, it is the custom to divide the mineral deposits into two groups—the metallic and the nonmetallic—a line of division which can hardly be defended except on the ground of long-established habit.

This book is the outcome of a desire to place the knowledge of mineral deposits on the broader and more comprehensive basis of a consistent genetic classification and thus bring it into a more worthy position as an important branch of geology. Opinions may differ as to whether our present knowledge is sufficient for such an undertaking. Believing that the time has come for a first attempt, I present this volume, in the hope that its shortcomings may be judged leniently.

The impetus to the work came during the preparation of a series of lectures a few years ago, and a course along the general lines followed in this volume has since then been presented annually at the Massachusetts Institute of Technology.

The general plan has been to select a few suitable examples to illustrate each genetic group of deposits. These examples have been chosen regardless of their geographic location, and it was of course necessary to give up any attempt to describe deposits in detail or to present all known examples of any particular type. As the larger part of my experience has been within the United States of America, a considerable number of examples were gathered from this country. This experience I owe to the United States Geological Survey, in which I have had the honor to serve for many years. My indebtedness to my friends and associates in that organization is greater than can be expressed in words.

Boston, *August*, 1913.

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CHAPTER I

INTRODUCTION

ECONOMIC GEOLOGY

The application of geology to the practical problems of the industries and the arts constitutes economic geology. This branch of the science includes as its most important division the study of deposits of useful minerals, but it also teaches the occurrence of underground waters, explains the derivation and constitution of soils in relation to agriculture, and applies geologic principles to the planning of important engineering works.

Only a part of the whole field of economic geology will be covered in these chapters. They will be confined to a description, by classes and type examples, of the occurrence, structure, and origin of the principal deposits of metallic and non-metallic minerals of economic importance. The subjects of coals, mineral oils, and structural materials could not be included without unduly increasing the bulk of the volume. Little space has been given to statistics, while the problems of correlation and origin have been treated rather fully. A general part describing principles of universal application precedes the detailed characterization of the various classes.

A complete treatment of the subject should include discussions of distribution, occurrence, structure, origin, production, and valuation of deposits, as well as statements of the uses of the materials mined, processes of mining and reduction, and criteria for judging the value of the products. Such a complete presentation will not be found in this volume, for it is believed that by examining the subject from a scientific rather than from a utilitarian viewpoint, the student will obtain a clearer insight into the geologic relationship of the various deposits.

Throughout its broad domain economic geology stands on the fundamental sciences of chemistry and physics. It is related on one side to theoretical geology, paleontology, mineralogy, and petrography; on the other side to mining, metallurgy, and many other technologic arts. A student who tries to approach the subject without the necessary knowledge of the allied sciences and arts is building on poor foundations. Even with this aid the study offers peculiar difficulties. The alteration of rocks close to many mineral deposits is intense and, as a result, the student who is familiar with only the fresh, unaltered specimens finds himself in the midst of puzzling and strange types that he is unable to classify with certainty. Altered andesites may assume the aspect of quartzites; a question may arise as to whether a silicified rock was once a limestone or a porphyry; diabases may at some places be converted into white fine-grained calcite-sericite-quartz rocks and at other places appear as aggregates consisting mainly of epidote and chlorite. These examples suffice to show that rock alteration is a subject of prime importance for the mining geologist.

MINERAL DEPOSITS

Definitions.—The outer shell of the globe is commonly called the earth's crust. Of this shell only the most superficial part is accessible. The radius of the earth is about 4,000 miles. The deepest shaft attains only about 6,000 feet,[1] the deepest bore-hole 7,350 feet. This crust consists of manifold mineral aggregates formed at different times and in various ways. Each individualized mass of mineral aggregates—such as a stratum, a lava flow, an intrusive mass of igneous rock, a dike, a vein, or a lenticular mass—is called a "formation," a "member," or in general a "geologic body." Geologic bodies which consist mainly of a single useful mineral—for instance, beds of pure gypsum or coal—or which contain, throughout or in places, valuable minerals that can be profitably extracted—for instance, veins containing disseminated gold—are called "mineral deposits." Geologic bodies that are not worked for any particular mineral or minerals, but for the aggregate of minerals—the rock itself—are usually designated as deposits of the particular

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[1] The gold mine of Morro Velho, Brazil, p. 190.

rock. Thus a mass of roofing slate is not spoken of as a mineral deposit, but as a slate deposit. Economic geology treats of the occurrence, composition, structure, and origin of those geologic bodies which can be technically utilized; it shows where they may be searched for and how their value may be ascertained.[1]

Technical Utility.—The limitation of technical utility must of course not be taken too literally, especially where questions of origin are concerned, for here, as in many other phases of the subject, applied geology merges into theoretic geology. Moreover, it is no uncommon occurrence that the useless of yesterday becomes the useful of to-day. Examples are easily cited. About 1900 the cupriferous monzonite of Bingham, Utah, which yields an average of 30 cents in gold and 14 cents in silver to the ton and 1.5 per cent. of copper, would probably not have been classed as an ore, but with modern methods of treatment it is an important ore of copper. The zinc minerals of the western States, valueless and even causing loss in the marketing of ores, can now be profitably sold. The tungsten ores of Colorado, thrown over the dump not long ago, have attained a value of $200 a ton. Low-grade gold ores—for instance, those of Mercur, Utah—considered as hopelessly refractory before 1890, became rich assets with the introduction of the cyanide process. Many iron ores rich in phosphorus were neglected until the Thomas process provided means for their profitable reduction. Monazite containing thorium acquired importance with the invention of the incandescent mantle for gas burners. New processes of reduction, the rising price of some commodity, inventions calling for rare and unused metals—any of these may suddenly cause a geologic body that has previously been valueless to become of great importance. Titanic iron ores form vast deposits which are now useless because of metallurgical difficulties but which some day will, no doubt, be utilized. This principle also works the other way. Decreasing prices may make a particular deposit unprofitable; that is what happened to many silver mines during the great decline in the price of silver which began in 1880. Great changes, mainly in the direction of rising prices have been brought about by the great war beginning in 1914. A large number of metals have doubled in price: They include silver, platinum, copper, lead, zinc, tin, antimony and

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[1] Stelzner and Bergeat, Die Erzlagerstätten, vol. 1, 1904, p. 1.

aluminum. Gold alone, being the standard by which other values are measured, remains stable.

Ore and Gangue.—These considerations bring us to the terms ore and gangue. "Ore" is a word which has been used in several meanings. An "ore mineral" is a mineral which may be used for the extraction of one or more metals. An "ore, "as the term is used here, is that part of a geologic body from which the metal or metals that it contains may be extracted profitably. Thus galena and malachite are ore minerals. An ore is practically always a mixture of minerals. Local usage has adopted several terms as substitutes for "ore." In the lead-zinc district of Missouri crude ore is called "dirt," while concentrates are called "ore." In Michigan the ore is called "rock" and the concentrates are termed "mineral." Gold-bearing gravels are not usually referred to as ore. The use of the term "ore" is not quite consistent. Ordinarily it implies a metal, but the expression "sulphur ore," meaning pyrite, is sometimes seen, and occasionally such terms as "sapphire ore" are found. The useless minerals occurring in the ore are termed "gangue." Thus, a gold ore may consist of quartz, calcite, siderite, native gold, auriferous pyrite, and galena. Here the first three are called "gangue minerals." The terms are not inflexible; for example, siderite may under some circumstances be utilized as an iron ore. Moreover, as stated above, what to-day is useless gangue may prove valuable ore to-morrow. It is therefore safe to make the definition of an ore rather wider than the present technical limits.[1]

It is hardly necessary to call attention to the differences in prices of metals which cause wide disparity in the amounts of different metals necessary to constitute ores. An iron ore must ordinarily contain at least 30 per cent. of iron—usually much more. A volcanic rock containing 15 per cent. of iron is far from being an iron ore, but quartz containing 0.05 per cent. of gold is a rich gold ore, worth $330 a metric ton; in fact, is little as 0.0001 per cent. of gold, equivalent to 1 gram to the metric ton, or a value of 66 cents a ton, if occurring in an ore with other useful substances, is ordinarily paid for by smelting works.

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[1] For a full discussion of the subject see J. F. Kemp, "What is an Ore?" *Jour.* Canadian Min. Inst., vol. 12, 1910, pp. 356–367. Also, *Min. and Sci. Press,* March 20, 1909.

DISTRIBUTION OF THE ELEMENTS

To obtain data regarding the relative distribution of the elements, several calculations have been undertaken on the basis of a great number of reliable rock analyses. Especially notable are the papers of Clarke,[1] Vogt,[2] and Washington.[3] Clarke used 1,200 analyses of American rocks; Washington 1,800 from various parts of the world. We are here chiefly concerned with the solid crust of the earth, although in passing it is deserving of notice that enormous quantities of salts are dissolved in the sea water, among them sodium chloride, sulphates of calcium, magnesium, and potassium, and carbonates of calcium and magnesium. The volume of salts in the sea water, according to Clarke, would be enough to cover the entire area of the United States (exclusive of Alaska) to a depth of 1.6 miles, or the whole globe with a stratum of sodium chloride 112 ft. deep. According to the same authority, the crust of the globe 10 miles thick, with an assumed average specific gravity of 2.5, contains about 93 per cent. of solid matter and 7 per cent. of sea water.

Composition of the Earth's Crust.—In calculating the average composition of the accessible portion of the solid crust it is necessary to consider the sedimentary and the igneous rocks. The sedimentary rocks form but a thin veneer compared with the igneous rocks. The average composition of the latter closely approximates that of the crust. Clarke calculates that the crust to a depth of 10 miles is composed of 95 per cent. of igneous rocks, 4 per cent. of shales, 0.75 per cent. of sandstones, and 0.25 per cent. of limestones. Van Hise and others arrive at somewhat different figures. The sediments average poorer in calcium, magnesium, and especially in sodium than the igneous rocks and thus show the effect of leaching. They also contain more potash and carbon dioxide, but on the whole they are similar in composition to the igneous rocks. According to Clarke the average of analyses of igneous rocks made in the laboratories of the United States Geological Survey is as follows:

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[1] F. W. Clarke, Geochemistry: *Bull.* 616, U. S. Geol. Survey, 1916, pp. 22–35. Many partial analyses are also included.

[2] J. H. L. Vogt, Ueber die relative Verbreitung der Elemente, etc.: *Zeitschr. prakt. Geol.*, 1898, pp. 225–238; 314–325.

[3] H. S. Washington, The distribution of the elements in igneous rocks: *Trans.* Am. Inst. Min. Eng., vol. 39, 1908, pp. 809–838.

AVERAGE ANALYSIS OF IGNEOUS ROCKS

O	47.29	S	0.103
Si	28.02	Cl	0.063
Al	7.96	F	0.10
Fe	4.56	Ba	0.092
Mg	2.29	Sr	0.033
Ca	3.47	Mn	0.078
Na	2.50	Ni	0.020
K	2.47	Cr	0.033
H	0.16	V	0.017
Ti	0.46	Li	0.004
Zr	0.017		
C	0.13	Total	100.000
P	0.13		

The eight elements first named above make up 98.56 per cent. of the igneous rocks.

Among the six principal metals shown in the average composition only iron, magnesium and aluminum are of economic importance. The lighter elements predominate, the atomic weight of each falling below 56 (Fe 55.9). In the average composition, the rarer metals titanium, zirconium, barium, strontium, manganese, nickel, chromium, and vanadium are represented, but except titanium, which amounts to 0.45 per cent., all these metals average below 0.1 per cent. Platinum, gold, silver, copper, lead, zinc, antimony, arsenic, tin, quicksilver, molybdenum, tungsten, and others are present in amounts less than 0.01 per cent.

For some of these more definite estimates have been made by Clarke and Steiger[1] from careful analyses of large, composite samples of rocks and clays. The average percentages are as follows: CuO, 0.0130; ZnO, 0.0049; PbO, 0.0022; As_2O_5, 0.0005. These figures considered as orders of magnitude have a high degree of probability; possibly they are a trifle too high. Silver may constitute 0.00001 and gold perhaps 0.0000005 per cent. of the crust.

The percentages of the useful metals given above do not by any means indicate the amount available for industrial use. That amount indeed is so infinitesimal in relation to the volume of the crust that it can not be expressed on the basis of percentages. The metals in the deposits of useful minerals then comprise only a minute fraction of the quantity of metals in the

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[1] *Jour.*, Washington Acad. of Sci., vol. 4, 1914, p. 57.

crust—a fraction which has been locally accumulated by this or that process of concentration.

In general igneous rocks contain more of the heavy metals than do the sedimentary rocks. We are well justified in regarding the former as the original source of these metals. Dissipation by solution accompanies sedimentation and the many metals found in traces in the sea water furnish evidence of this. On the other hand, it is true that certain kinds of sedimentation will result in a concentration of metals, such as iron, zinc, cobalt, nickel and vanadium.

Vogt and Washington have also formulated some rules concerning the relationship of certain metals with certain rocks. It is obvious, however, that a distinction should be made as to whether the metal is an integral part of the rock or whether simply deposits of the metal occur in the rock. Thus, for instance, lead deposits are characteristic of many limestones, but it may be doubted whether lead is a primary constituent of limestone; it is present because the rock had the power of precipitating the metal from its solution.

In highly siliceous rocks, especially in granites and in the pegmatite dikes accompanying them, we find minerals containing fluorine, boron, lithium, zirconium, tin, tungsten, tantalum, molybdenum, thorium, and beryllium. Highly sodic magmas are also accompanied by a great number of rare metals.

On the other hand, basic rocks in which the darker constituents predominate contain phosphorus, sulphur, chlorine, copper, chromium, nickel, cobalt, titanium and vanadium, and some of them, chiefly peridotites, contain platinum and diamonds. Gold and silver exist in minute quantities in many rocks, particularly in those of acidic types, like granite and rhyolite.

These rarer metals are not everywhere present in similar rocks. Platinum, for instance, is contained in the peridotites of the Ural Mountains, but the peridotites of the Sierra Nevada are poor in that metal and the similar rocks in the Coast Ranges of California contain little platinum, as do the serpentines of Asia Minor and of Italy. Similar conditions characterize the occurrence of rarer metals in granitic rocks.

TRACES OF METALS IN ROCKS

General Statement.—In order to formulate a theory or a hypothesis of the origin of mineral deposits it is most desirable to ascertain to what extent the different rocks contain the rarer metals.

J. G. Forchhammer and L. Dieulafait began examinations for this purpose about 1860 and found traces of silver, copper, lead, bismuth, nickel, cobalt, zinc, arsenic, antimony, and tin in many rocks. Somewhat later F. von Sandberger followed up this line of investigation and ascertained that the dark silicates of many rocks contained lead, copper, tin, antimony, arsenic, nickel, cobalt, bismuth, and silver. Some doubt has been expressed as to a few of these results and it is believed that in some of Sandberger's specimens the metals were derived from adjacent veins or from the reagents or the vessels used in the analyses. However, in spite of analytical difficulties, the presence of many of those metals in various igneous and metamorphic rocks is clearly proved. Many ordinary analyses show the presence of chromium, cobalt, and nickel in basic rocks like peridotites, serpentines, and pyroxenites. Some of these rocks contain as much as 0.76 per cent. of Cr_2O_3 and up to 0.3 per cent. of $(Ni,Co)O$. Traces of nickel and cobalt are often found in diabases, gabbros, and basalts; occasionally in diorites. A little vanadium is common in all rocks—usually only 0.01 to 0.05 per cent. of V_2O_3.

In the following paragraphs some of the most reliable data regarding traces of rarer metals are compiled. More extensive references will be found in Clarke's "Data of geochemistry."

Copper.—A. C. Lane[1] states that the Keweenawan "traps" average 0.02 per cent. of copper. F. F. Grout[2] found 0.029 and 0.02 per cent. in fresh specimens of the same series from Minnesota. J. Volney Lewis[3] says that some of the New Jersey diabases or "traps" contain chalcopyrite and that copper is also present in the pyroxene of these rocks. The average, according to numerous analyses, is 0.025 per cent. of CuO. R. C. Wells found 0.03 per cent. of copper in a perfectly fresh basaltic lava from The Dalles, Oregon.

Analyses made for W. H. Weed in the laboratory of the United States Geological Survey show that Butte granite or quartz monzonite from a quarry near Walkerville, Montana, contains 0.006 per cent. of copper. The quartz and feldspar, forming

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[1] Mine waters, *Proc.*, Lake Superior Min. Inst., June, 1908, p. 86.

[2] *Econ. Geol.*, vol. 5, 1910, p. 471.

[3] *Econ. Geol.*, vol. 2, 1907, pp. 242–257.

91 per cent. of the rock, contain little or no copper; the mica and hornblende, which constitute 7 per cent. of the rock, yield 0.047 per cent. of copper; there is thus nearly eight times as much copper in the ferromagnesian minerals as in the rock. The altered rock surrounding the veins carries more copper than the fresh rock.

E. T. Allen examined 18 samples of fresh gabbros and diorites from the Encampment district, Wyoming, and found copper in all, the largest quantity noted being 0.02 per cent. of CuO.[1]

A composite sample of seventy-one Hawaiian lavas yielded Geo. Steiger 0.0155 per cent. of copper. H. I. Jenssen found 0.034 per cent. of copper in an andesite from Fiji.[2] E. Comanducci reported 0.0854 per cent. of CuO and 0.0038 per cent. of CoO in volcanic ash from Vesuvius.[3]

J. B. Harrison[4] examined 36 igneous and metamorphic rocks from British Guiana and found that 6 contained no copper and 12 contained copper in traces only; those carrying most copper were diabases and porphyrites; a feldspathic tuff yielded 0.13 per cent. The average for the series was 0.025 per cent. of copper. In some of these rocks the copper may have been contained in secondary disseminated sulphides.

In a fresh granodiorite from Steamboat Springs, Nevada, W. H. Melville[5] detected copper, lead, arsenic, and antimony. J. D. Robertson[6] found from 0.0024 to 0.0104 per cent. of copper in granite, porphyry, and diabase from the Archean of St. Francis Mountain, in Missouri. The average was 0.006 per cent. Lead and zinc were also recognized. The adjacent Silurian and Carboniferous limestones also contained these metals, but in smaller quantities.

Native copper in minute scales is rather common in shales and copper sulphides, contemporaneous with the metamorphism, occur in many amphibolites. Copper has been repeatedly detected in sea water and is contained in the red and blue mud dredged from the deep seas.

[1] A. C. Spencer, *Prof. Paper* 25, U. S. Geol. Survey, 1904, p. 49.

[2] *Chem. News*, vol. 96, 1907, p. 245.

[3] *Gazz.* chim. ital., vol. 36, pt. 2, 1906, p. 797.

[4] Report on the petrography of the Cuyuni and Mazaruni districts. Georgetown, Demerara, 1905.

[5] *Mon.* 13, U. S. Geol. Survey, 1888, p. 350.

[6] Missouri Geol. Survey, vol. 7, 1894, pp. 479–481.

From this evidence the conclusion may be drawn that probably all igneous rocks contain appreciable amounts of copper and that acidic rocks contain less than basic rocks. The copper is largely associated with the ferromagnesian silicates, and in the lavas at least it appears to be present as a silicate.

Lead and Zinc.—In analyzing the quartzose porphyries of Leadville, Colorado, supposed to be free from sulphides, W. F. Hillebrand[1] found that of 18 carefully selected samples 15 contained lead, the richest carrying 0.0064 per cent. of PbO; the average was 0.002 per cent. The same analyst found 0.008 and 0.0043 per cent. of ZnO in similar rocks.

J. D. Robertson[2] determined an average of 0.004 per cent. of lead and 0.009 per cent. of zinc in the Archean rocks from Missouri mentioned above. J. B. Weems[3] determined lead and zinc in the limestones and dolomites of the Dubuque region, Iowa. The average of 9 samples gave 0.00326 per cent. of lead and 0.00029 per cent. of zinc. J. B. Harrison looked for lead in 23 samples of rocks from British Guiana and was able to determine the metal in five; the maximum obtained was 0.02 per cent. L. Dieulafait detected zinc in hundreds of samples of Jurassic limestone from central France.

On the other hand, W. F. Hillebrand[4] was unable to detect lead or zinc in samples of limestone from Mexico, near important lead deposits. Henry W. Nichols[5] found no lead, copper, or zinc in calcareous concretionary deposits of the Challenger Banks, near Bermuda. Zinc is reported by Dieulafait in sea water and in ashes of sea weeds. Lead apparently does not exist in the water of the ocean.

Gold and Silver.—A large number of experiments have been undertaken to decide the question whether fresh igneous or sedimentary rocks contain gold and silver. In mining districts where the solution of this problem has been frequently attempted it is difficult to obtain perfectly satisfactory samples, free from contamination by circulating water. Furthermore, contamination is possible from fluxes, from the dust of assay rooms, from mortars, or from bucking boards. The mere statement that the assay

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[1] *Mon.* 12, U. S. Geol. Survey, 1886, pp. 591–594.

[2] Missouri Geol. Survey, vol. 7, 1894, pp. 479–481.

[3] Iowa Geol. Survey, vol. 10, 1900, p. 566.

[4] Oral information.

[5] *Econ. Geol.*, vol. 2, 1907, p. 309.

indicates gold and silver in a rock is not sufficient. It must be corroborated by a statement of the methods used and accompanied by the evidence of microscopic examination as to the freshness of the rock.

It is satisfactorily proved that many fresh, massive igneous rocks contain gold. The best evidence thus far brought forward is probably that afforded by the granite from the Altar district, Sonora, Mexico, described by G. P. Merrill.[1] The gold occurs embedded in fresh quartz and feldspar. W. Möricke[2] found gold in a pitchstone from Chile and believed the metal to be primary. Gold was found by W. F. Ferrier in a fresh syenite from Kamloops, British Columbia. According to R. W. Brock[3] probably primary gold was found in a porphyry dike on North Fork of Salmon River, West Kootenai, British Columbia. Of twelve dikes of porphyry at different points in West Kootenai, six contained gold, most of them being wholly unaltered. Brock also states that a sample of alkali syenite porphyry in the Valkyr Mountains, east of Lower Arrow Lake, British Columbia, contained gold that was visible to the naked eye.

Many of the statements of this sort in the literature must be critically scanned, for it is not uncommon to find gold deposited by mineralizing solutions in massive rocks, especially in schists, under circumstances closely simulating original deposition. Statements regarding primary gold in the chloritic schists of the Sierra Nevada refer to occurrences of this class.

It is frequently said that gold occurs in pegmatite, but few of the assertions have the requisite backing of complete evidence. The probability is strong, however, that gold is present in such dikes. One of the most definite descriptions of this mode of occurrence is furnished by J. Catharinet.[4] Sperrylite, an arsenide of platinum, is stated by Catharinet to occur with the gold. Another case is reported by C. De Kalb[5] from Mohave, California. References in the literature to primary gold in rocks from the Ural Mountains and from Australia do not appear to be sufficiently substantiated.

Having procured samples of various rocks, most of them far

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[1] *Am. Jour. Sci.*, 4th ser., vol. 1, 1896, p. 309.

[2] *Min. pet. Mitt.*, vol. 12, 1891, p. 195.

[3] *Eng. and Min. Jour.*, vol. 77, March 31, 1904.

[4] *Eng. and Min. Jour.*, vol. 79, 1905, p. 127.

[5] *Trans.*, Am. Inst. Min. Eng., vol. 38, 1908, p. 312.

from mining districts, Luther Wagoner,[1] of San Francisco, assayed them with the results as given in the accompanying table. His method consisted in cyanide treatment of 40 or 50 grams of material followed by blowpipe cupellation. All particulars of the operations are detailed.

Wagoner found that the purest obtainable reagents, such as soda, borax, and cyanide of potassium, contain gold and silver. A sample of Merck's "C. P." carbonate of soda contained 3 grams of silver to the ton. A sample of cyanide of potassium yielded 147 milligrams of gold and 26.05 milligrams of silver to the metric ton. Wagoner's results merit attention, but it would be desirable to have them checked. His figures for silver seem high. It will be noted that gold values obtained by him are as low as 5 milligrams, or 1/13 of a grain, to the ton. J. R. Don, using the fire assay, was able to detect amounts as low as 6.4 milligrams per metric ton.

GOLD AND SILVER CONTENTS OF VARIOUS ROCKS
[Milligrams per metric ton]

Rock and locality	Gold	Silver
Granite, Lake Tenaya, California	104	7,660
Granite, Lake Tenaya, California	137	1,220
Granite, head of American River, California	115	940
Syenite, Candelaria, Nevada	720	15,430
Granite, Candelaria, Nevada	1,130	5,590
Sandstone, Colusa County, California	39	540
Sandstone, Angel Island, California	24	450
Sandstone, Russian Hill, San Francisco, California	21	320
Basalt, Petaluma, California	26	547
Diabase, Mariposa County, California	76	7,440
Marble, Tuolumne, California	5	212
Marble, Carrara, Italy	8.63	201

Sea water is known to contain small quantities of gold and silver. To illustrate the difficulty and uncertainty attending the measurement of minute traces of metals the following results, reached by several chemists, are given:

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[1] *Trans.*, Am. Inst. Min. Eng., vol. 31, 1901, pp. 798–810.

GOLD AND SILVER IN SEA WATER
[Milligrams per metric ton]

Chemist	Gold	Silver
Münster (Norway)	5–6	19–20
Liversidge (Australia)	33–65	65–130
Don (New Zealand)	4.5	None.
Wagoner (San Francisco)	12–16	1500–1900
Malaguti and Durocher	None.	9

Wagoner found 457 milligrams of gold and 54.4 grams of silver to the metric ton of salt evaporated from sea water, and later reported both metals in appreciable quantities in deep-sea dredgings.[1]

J. R. Don[2] assayed a large number of rocks both close to and at a distance from ore deposits, to ascertain their content of gold and silver. His work indicated in general that gold and silver are contained only in rocks that have been impregnated with pyrite in the vicinity of ore deposits. These results, however, do not agree with those of other analysts and are especially contradicted by Wagoner's experiments. Don's conclusions, broadly speaking, are probably true, but his methods largely excluded detection of the possible occurrence of gold in the silicates of the rocks.

A few evidently authentic occurrences of primary gold in crystalline schists are known. Spurr[3] describes an auriferous quartz diorite gneiss from the Ayrshire mine, Lomagundi, Mashonaland, Africa. The gneiss forms a dike 20 feet in width, of which 15 feet is mined; the average gold content is said to be about $15 a ton. The gold is intergrown with orthoclase, plagioclase, quartz, epidote, and hornblende; it is especially associated with the bands of hornblende and is, according to Spurr, unquestionably of primary origin—that is, it crystallized during the metamorphism with the other constituents of the gneiss. This does not, however, absolutely prove that it was a

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[1] *Trans.*, Am. Inst. Min. Eng., vol. 38, 1907, p. 704.

[2] The genesis of certain auriferous lodes: *Trans.*, Am. Inst. Min. Eng., vol. 27, 1898, p. 564.

[3] J. E. Spurr, Native gold original in metamorphic gneisses: *Eng. and Min. Jour.*, vol. 76, 1903, p. 500.

constituent of the igneous rock from which the gneiss was derived.

Lacroix[1] reports gold in extremely abundant and minute crystals in a biotite gneiss from Madagascar; also in a magnetite-bearing quartzite from the same island. F. C. Lincoln has recently given a comprehensive review of the data regarding gold in rocks.[2]

Neither primary native silver nor visible silver compounds are reported to occur in igneous or metamorphic rocks. W. F. Hillebrand found silver in a number of porphyries from Leadville, Colorado, that were collected and assayed with the greatest care; the average amount was 0.0265 ounce to the ton (884 milligrams to the metric ton).[3] Traces of gold were rarely found.

J. W. Mallet[4] found silver in two samples of volcanic ash from the Andes, to the amount of about 10 grams to the metric ton.

TENOR OF ORES

While it is not possible to give exact data as to the minimum values which ores of the different metals should have for profitable extraction, some approximate statements may be useful.[5] Local conditions, price of metals, the nature of the ores, and the association of the metals must of course be considered.

Iron.—Iron ores from the Lake Superior region usually contain 50 to 60 per cent. of iron; but iron ores which contain less than this may be utilized, especially where other conditions are favorable. The Clinton ores of Alabama contain as little as 30 per cent. of metallic iron; some types of easily concentrated magnetites may contain as low as 25 per cent. and still yield a profit.

Copper.—Copper ores of the Lake Superior region may be treated with profit, under favorable circumstances, with as little

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[1] Lacroix, A., Sur l'origine de l'or de Madagascar: *Compt. Rend.*, vol. 132, January 21, 1901, pp. 180–182.

[2] *Econ. Geol.*, vol. 6, 1911, pp. 247–302.

[3] *Mon.* 12, U. S. Geol. Survey, 1886, pp. 591–594.

[4] *Chem. News*, vol. 55, 1887, p. 17.

[5] J. F. Kemp, Problem of the metalliferous veins: *Econ. Geol.*, vol. 1, 1906, pp. 207–232.

as 0.5 per cent. of metallic copper, though they ordinarily average somewhat higher. Sulphide copper ore of the usual type can rarely be utilized if it contains below 1.5 per cent. of copper unless gold and silver are present also, and in many districts the ores must average considerably higher than this. The ores treated at Ely, Nevada, contain about 2 per cent. copper, besides gold and silver to the value of 40 cents a ton.

Lead.—In northern Idaho lead ores which contain 5 to 6 per cent. of lead and 3 ounces in silver to the ton are profitably mined. Non-argentiferous ores which assay from 5 to 7 per cent. of lead are utilized in southeastern Missouri.

Zinc.—Zinc ores vary considerably according to locality. At Joplin, Missouri, much of the crude material hoisted yields less than 3 per cent. of zinc sulphide and a little lead. This is concentrated to about 60 per cent. of zinc. In localities more remote from markets, as in Colorado, Utah, and Idaho, only high-grade zinc ores can be profitably treated or shipped.

Silver.—With rising metal prices pure silver ores again attract attention. Quartzose ores should contain not less than 15 ounces to the ton. The usual ores contain silver in association with lead, copper, or gold or with all three. In complex ores smelters rarely pay for less than 2 ounces of silver and 0.01 ounce of gold to the ton. Gold and silver are separated from the lead or copper bullion by zinc desilverization or electrolytic refining and the cost of that process, of course, imposes the necessity of a certain minimum tenor of gold and silver for profitable extraction, but at many plants gold and silver, although present in less than these small quantities, are obtained as by-products through the necessity of eliminating some objectionable constituent, like arsenic, from the bullion.

Gold.—Gold has been profitably extracted from ores yielding less than one dollar to the ton, but the ordinary gold quartz ores—for instance, those of California—yield about $5 to the ton; those of Nevada, Colorado, and some other States usually contain more. On a large scale gold ores containing from $2.50 to $3 a ton, or even less, may be worked, as at the Treadwell mines, in Alaska. In gold gravels worked by the hydraulic process as little as 4 or 5 cents to the cubic yard may be profitable. By dredging, gravels containing 8 to 15 cents a cubic yard may be utilized in California; in Alaska they should contain from 50 cents to $1 a cubic yard. In the last few years the costs of gold dredging have been brought down to about 4 cents a cubic yard.

Tin, Etc.—Tin ores range from 1.5 to 3 per cent. in tin, but in tin-bearing gravels a much smaller tenor is sufficient to yield a profit. Ores of quicksilver contain at least 0.3 per cent. of that metal; aluminum ores at least 30 per cent. of aluminum. Nickel should be present to the amount of 2 per cent. or more to constitute a workable nickel ore. Manganese ore should contain 50 per cent. of that metal, but less is required if iron is also present. Chromium ore must contain about 50 per cent. of chromic oxide. Poorer ores may be used if amenable to concentration.

PRICE OF METALS

The prices which the various metals bring express the result of their abundance, of the demand for them, and of the cost of reduction of their ores. The value of gold is fixed by international agreement, hence it constitutes the standard by which the prices of all other commodities are measured. Aluminum, the most common of all metals, brings a high price because it can be produced from only a few of the minerals containing it.

In the following table the first column represents what may be called the "normal" prices for metals. Even in normal times there is, of course, constant fluctuations and some metals like copper, iron and tin are especially susceptible to economic influences. The great World War beginning in 1914 proved to have a potent influence on prices and by consulting the second column it will be found that the prices of most metals had been doubled in 1918. With the re-establishing of normal conditions the prices, except for gold and silver, are likely to decline.

Many other substances used for war materials have reached an abnormal price in 1917: High-grade manganese ore sells at $1 per unit and about the same price is obtained for chromite. Tungsten ore brings $20 to $26 per unit of WO_3, and molybdenite over $2 per pound.

The values of the generally used metals in 1914 and 1918 compare as follows:

COMPARATIVE VALUES OF METALS

	March, 1914	February, 1918
Platinum	$44.00 per ounce	$106.00 per ounce
Gold	20.67 " "	20.67 " "
Silver	0.57 " "	0.87 " "
Quicksilver	0.52 per pound	1.71 per pound
Nickel	0.45 " "	0.50 " "
Tin	0.37 " "	0.85 " "
Aluminum	0.19 " "	0.37 " "
Copper	0.14 " "	0.23 " "
Antimony	0.07 " "	0.14 " "
Zinc	0.05 " "	0.08 " "
Lead	0.04 " "	0.07 " "
Pig iron	0.006 " "	0.015 " "

Many interesting data on the total quantities of metals produced in the world and on the largest amounts mined in any one deposit are given by J. H. L. Vogt.[1]

PRODUCTION OF ORE AND METAL

An interesting table reducing metal production in the United States to a uniform basis of short tons is published annually in Mineral Resources.[2] From this it is seen that in 1915, for instance, were produced nearly 33,000,000 tons of pig iron, nearly 700,000 tons of copper, 507,000 tons of lead, 458,000 tons of zinc, 2,570 tons of silver and 167 tons of gold.

The ore production in short tons of the same year was as follows: Iron ores, 62,150,000; copper ore, 43,500,000; lead ore, 7,500,000; zinc ore, 18,200,000; silver ore, 1,400,000; gold ore, 11,300,000, all in round figures.

WEIGHTS AND MEASURES

Before leaving this part of the subject a few words on weights and measures may be added. The contents of base-metal ores, such as iron, lead, zinc, and copper, are measured by percentage.

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[1] Beyschlag, Krusch and Vogt, Die Lagerstätten, etc., vol. 1, 1909, pp. 187–200.

[2] J. P. Dunlop, Metals and Ores in 1914 and 1915; *Mineral Resources*, U. S. Geol. Survey, pt. 1, 1915.

For lead and copper the figures given often do not mean the exact content by wet analysis, but by the dry assay, which is 1½ per cent. or more lower than the exact content. In some cases the lead is determined by wet assay of the button from a crucible assay, which places the percentage obtained still farther below the actual content. The smelter pays for the metals by the "unit," which means 1 per cent., or 20 pounds to the ton, or else by a "basis price" for a given percentage of metal, say 55 per cent. for bessemer iron ores or 65 per cent. for Joplin zinc concentrates. Tungsten ore is sold per unit of tungsten trioxide for ore carrying 60 per cent. or more of this compound. Deductions and allowances based on the presence or absence of certain elements and certain other rules complicate the smelter schedules.[1]

Precious metals in ores are measured in England and its colonies by troy fine ounces and pennyweights, per long or short ton. In the United States decimal fractions are substituted for pennyweights; gold is often reported in dollars and cents, $1 corresponding closely to 1 pennyweight. Silver is measured in fine ounces, the pennyweights always being omitted. The short ton is always used. Practically all other nations measure these metals in grams per metric ton, a far more sensible way. For comparison the following data, computed and arranged by W. J. Sharwood,[2] are given:

Conversion Tables.—The gram is taken as 15.4320 grains. The value of a troy ounce of fine gold is assumed as being exactly $20.67, instead of $20.6718346+,[3] resulting in an error of less than 1 in 10,000. Values in English coin are based on the assumption that an ounce of fine gold is worth 4.25 pounds sterling, or 85 shillings, or 1,020 pence; this is too high by about 1 part in 2,000, the true value being 1,019.45 pence. It is useless to attempt a closer approximation in practical work, for the simple reason that gold bullion assays are rarely reported closer than the nearest half millieme, or to within 1 part in 2,000. At the values adopted one dollar is equivalent to 4.11224 shillings, and one pound sterling to $4.86353.

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[1] C. H. Fulton, The buying and selling of ores and metallurgical products, *Tech. Paper* 83, U. S. Bureau of Mines, 1915.

[2] Conversion tables for assay valuations, *Mines and Minerals*, January, 1909, p. 250.

[3] The United States Mint Bureau and the United States Geological Survey use tables compiled on the basis of $20.671834625323.

VOLUME AND WEIGHT OF FINE GOLD AND SILVER

	One cubic centimeter	One cubic inch	One cubic foot
Fine silver:			
Weight: grams	10.57	173.21	299307.00
Weight: troy ounces	.339825	5.5687	9622.72
Fine gold:			
Weight: grams	19.3	316.269	546,513
Weight: troy ounces	.6205	10.1680	17,570.39
Value: United States dollars	$\$12.82^{57}$	\$210.17	\$363,180
Value: pounds sterling	£2.647	£43.214	£74,674

ASSAY VALUATIONS

Values	One part in	Per cent.	Per metric ton			Per long ton of 2,240 pounds						Per short ton of 2,000 pounds	
			Troy ounces	United States dollars (gold)	Grams	Troy ounces	Oz.	Dwt.	Gr.	United States dollars (gold)	Troy ounces	United States dollars (gold)
One per cent.	100	1	321.50	$6,645.40^{6}$	10,000	326.666	326	13	8	6,752.20	291.666	6,028.75
One gram per metric ton = one part per million.	1,000,000	0.0001	0.03215	0.66^{45}	1	0.03266		. . . . . .	15.68	0.67^{5}	0.029166	0.60^{29}
One troy ounce per short ton.	29,166.66	.00342857 $=\frac{24}{7000}$	1.1023	22.78^{45}	34.2857	1.120	1	2	9.6	23.15	1	20.67
One troy ounce per long ton.	32,666.66	.0030612 $=\frac{3}{980}$	0.9842	20.34^{34}	30.612	1	1	. . . . . .	. . . . .	20.67	0.892857	18.45^{8}
One troy ounce per metric ton.	32,150	.0031104 $=\frac{2}{643}$	1	20.67	31.104	1.016	1	. . . . . .	7.7	21.00	0.90720	18.75^{2}
One dollar gold per short ton.	602,875	.00016587	.0533285	1.1023	1.6587	.054185	. . . .	1	2	1.12	0.048379	1

CONVERSION TABLES—WEIGHTS

	Grains	Penny-weights	Troy ounces	Avoirdupois ounces	Avoirdupois pounds	Grams	Fine gold value	
							United States	British
One grain	1	0.041666	0.0020833	0.00228571	0.000142857	0.0648	4.306 cts.	2.125 pence
One pennyweight	24	1	0.0500	0.0548571	0.00342857	1.5552	$1.0335	4.25 shillings
One troy ounce	480	20	1	1.0971428	0.0685714	31.104	$20.67	85 shillings
One troy pound	5,760	240	12	13.165714	0.822857	373.248	$248.04	£51
One avoirdupois ounce	437.50	18.22917	0.911458	1	0.06250	28.35	$18.84	77.474 shillings
One avoirdupois pound	7,000	291.666	14.58333	16	1	453.60	$301.4375	£61.97
One milligram	0.015432	0.000643	0.00003215	0.000035274	0.0000022046	0.0010	0.06645 ct.	0.033 penny
One gram	15.432	0.643	0.03215	0.035274	0.0022046	1	66.45 cts.	2.73275 shillings
One kilogram	15,432	643	32.15	35.274	2.2046	1,000	[1]$664.54	£136.64

The price of silver in the United States is stated in cents per troy ounce 1,000 fine; in England in pence per troy ounce of sterling silver 925 fine: 1 cent per fine ounce = 0.457 pence per ounce 925 fine; 1 penny per ounce 925 fine = 2.19 cents per fine ounce. With silver at 50 cents per ounce, 3 grams per ton equals about 5 cents United States currency.

Precious stones are measured by the carat. One carat (diamond weight) is equal to 3.2 grains troy, or 206 milligrams. The "metric carat," now universally used, is equal to 200 milligrams.

[1] The United States Mint Bureau uses $664.60.

CHAPTER II

THE FORMATION OF MINERALS

SOLUTION AND PRECIPITATION

General Features.—A mineral is an inorganic body of definite chemical composition found in the crust of the earth. Solid minerals[1] are formed by changes in chemical energy in systems which contain one fluid or vapor phase. Their development, therefore, generally indicates a transition from a mobile to a less mobile form of matter. In the great majority of cases the minerals in nature are formed by precipitation from solutions,[2] and this, therefore, becomes a process of the highest importance for mineral genesis.

Supersaturation and consequently precipitation are controlled by the thermodynamic environment of the system. Temperature and pressure are the most important agencies though at times electric energy and light may also be active.

In most cases the formation of minerals involves a chemical reaction subject to pressure and temperature, and this may be brought about in several ways:

1. By reactions between liquids or liquid solutions.
2. By reactions between gases or gaseous solutions.
3. By reactions between liquids or liquid solutions and gases.
4. By reactions between solids and liquids or liquid solutions or gases.

Few minerals are formed below the freezing point of water.

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[1] The liquid minerals are practically confined to water, mercury and to certain hydrocarbons. The latter, as well as certain undercooled liquids like obsidian, are not, however, considered as minerals as they have no definite composition expressible in a chemical formula.

[2] Certain chemical reactions involving the formation of new minerals may apparently take place by the interaction of solids under the influence of heat and pressure, but it is perhaps probable that here too a vapor phase or a fluid phase is present in the system.

Their upper limit of development is marked by the temperature at which they become unstable or melt.

After the inception of mineral formation concentration to larger masses or deposits may be effected by the continuation of similar processes or reactions. Quite frequently, however, there are other causes which have contributed: Gravity may cause the sinking of heavy minerals in lighter media as when anhydrite crystals sink to the bottom in sea water, or when magnetite sinks in a residual rock melt. Solution and disintegration may do its share, as when native gold is set free from a quartz matrix, and is separated by gravity into workable gold deposits; or, as when, in a rock composed of calcite and calcium phosphate, the former is dissolved, leaving a phosphate deposit of economic value.

Influence of Pressure.—The effect of change of pressure, according to LeChatelier's law, is as follows: When the pressure in a system in equilibrium is increased that reaction takes place which is accompanied by a diminution in volume; and when the pressure is diminished a reaction ensues which is accompanied by an increase in volume. "The direction in which change of concentration will occur with change of pressure can be predicted, if it is known whether solution is accompanied by increase or diminution of the total volume. If diminution of the total volume of the system occurs on solution, as is the usual case, increase of pressure will increase the solubility; in the reverse case increase of pressure will diminish the solubility."[1] In general, a decrease of pressure, which results when solutions ascend in the earth's crust, will be favorable to precipitation. The influence of pressure is, however, in most cases slight. For instance, the solubility of sodium chloride (in grams of salt in 1 gram of solution) at the pressure of one atmosphere is expressed by 0.264 and at 500 atmospheres by 0.270. In systems in which one or more of the components are volatile the effect of pressure may be very great; carbon dioxide, for instance, held in water by pressure, may increase the solubility of calcium carbonate owing to the formation of bicarbonate.

The effects of uniform, or hydrostatic, pressure are much less marked than are the effects of stress, or unequal pressure. Under conditions of stress a given pressure will lower the melting

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[1] Alexander Findlay: The Phase Rule, 1908, p. 107.

temperature far more rapidly than when the pressure is equal from all sides.[1]

Influence of Temperature.—In a solution of various salts in water or in a silicate melt changes in temperature are far more effective in producing precipitation than changes in pressure. Van't Hoff's law states: When the temperature of a system in equilibrium is raised that reaction takes place which is accompanied by absorption of heat; and, conversely, when the temperature is lowered that reaction occurs which is accompanied by an evolution of heat.[2] In the great majority of cases increase of temperature promotes the solubility of salts, and decreasing temperature—say by the cooling of ascending thermal waters or of magmas—promotes precipitation. The common rule for salts—to which a number of exceptions may be noted—is that the solubility increases to temperatures of 75° C. or 150° C. beyond which a lessening of the quantity dissolved may often be noted. Breaks in the solubility curve usually indicate the limit of stability for a particular salt, less hydrated forms, for instance, coming in at higher temperatures.

In any hot, complex solution, occurring in nature, decreasing temperature will, in general, cause precipitation of some mineral; with continued cooling a series of other minerals may be precipitated, as the solubility limit of each is reached.

As no compounds are absolutely stable under the varying conditions obtaining in the crust, it follows that minerals once formed may subsequently be brought into solution, transported, redeposited, or indeed wholly decomposed so that its elements may enter into new combinations.

Precipitation by Evaporation of the Solvent.—The salts contained in a solution are naturally precipitated when evaporation at the surface so reduces the amount of the solvent that supersaturation ensues. The deposits of gypsum and salt in various formations are familiar results of this process. In some cases carbon dioxide or other gases may be the solvent; the precipitation of calcium carbonate follows, for instance, in springs at

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[1] John Johnston and Paul Niggli, The general principles underlying metamorphic processes. *Jour. Geology*, vol. 21, 1913, pp. 481–516 and 588–624.

John Johnston, Pressure as a factor in the formation of rocks and minerals. *Jour. Geology*, vol. 23, 1915, pp. 730–747.

[2] Alexander Findlay, The phase rule, 1908, p. 68.

their point of issue when the carbon dioxide escapes which holds the salt in solution as a bicarbonate.

Precipitation by Reaction between Solutions.—Mingling of different solutions is one of the most common occurrences in nature, as when rivers discharge their waters into the sea or as when ascending hot waters meet surface waters of different composition. Precipitation of chemical compounds results when any combination of the various ions in the solution can form to a sufficient extent to be insoluble in the liquid present. Solutions in nature are usually complex and the various reactions are more or less interfered with. In general, according to Nernst's law, the solubility of a given salt is reduced by the presence in the solution of another salt which has a common ion but is increased by the presence of another salt with no common ion. For instance, the solubility of lead chloride is decreased by the presence of the chloride of calcium or magnesium. The presence of alkaline carbonates decreases the solubility of $FeCO_3$; the solubility of $NaCl$ is decreased by $CaCl_2$; while the solubility of $CaSO_4$ (gypsum) is increased in a $NaCl$ solution. If calcite is treated with a saturated solution of $FeCO_3$, $ZnCO_3$, or $MgCO_3$, a part of the calcite will be dissolved while a corresponding part of the other carbonates is precipitated; the solubility of $CaCO_3$ in water is increased by Na_2SO_4 or $NaCl$ but decreased by $MgCO_3$.

In mixed solutions precipitation is often delayed, as shown, for instance, by the slow precipitation of barite ($BaSO_4$) due to the presence of sodium and magnesium chlorides in certain mine waters consisting of salt brines. Nernst's law offers an explanation of these anomalies.

Slow precipitation in dilute solutions generally results in large crystals being formed, while rapid precipitation results in colloids or fine powders.

Precipitation by Reactions between Aqueous Solutions and Solids.—In nature, solutions act constantly upon solid minerals. At the surface all rocks and mineral deposits are moistened by rain water which also may descend to great depths in porous, fissured or broken material. Rising hot waters soak into the rocks from the fractures upon which they ascend. Minerals are attacked to a greater or lesser degree by these various kinds of waters; they are decomposed and partly or wholly go into solutions. From these solutions new minerals are deposited in open spaces and this is a very common mode of mineral formation. But the changes also proceed in the solid rocks themselves and such processes by which new minerals may take the place of old ones are called metasomatism or replacement. The water penetrates the rocks in capillary openings. By the phenomenon known as adsorption, the film of liquid on the solid contains more than an average amount of material in solution and these films are likely to become supersaturated in advance of the remainder of the solution, so that chemical reactions will be facilitated. In this manner the mineralogical and structural character of rocks may be changed: Chlorite may replace augite, and sericite and quartz may replace feldspars. Metallic ores are often formed by replacement. Limestone may, for instance, be permeated by a solution of zinc sulphate with the result that the calcium carbonate is replaced by zinc carbonate with faithful preservation of the limestone structure, while calcium sulphate is carried off in solution. It is not even necessary that the replacing mineral should have an element in common with the older mineral. Pyrite or galena may replace feldspars or calcite grains. The replacing mineral may even develop as perfect crystals in the older mineral. The phenomena of replacement are of the utmost importance for the genesis of mineral deposits.

Precipitation by Reactions between Gases or between Gases and Solutions.—Gases may produce precipitation in solutions. Hydrogen sulphide in some mine waters precipitates cuprous sulphide from cupric sulphate. Less important is the action between gases: Native sulphur may be precipitated in volcanic regions by a mixture of hydrogen sulphide and sulphur dioxide.

Crystalline Minerals.—The minerals may be precipitated as crystalloids or as colloids. In mineral deposits formed in depth and at temperatures higher than those prevailing at the surface crystalloids are almost exclusively present, as they also are in igneous and metamorphic rocks. The molecules have been allowed to arrange themselves in the symmetry of one of the six crystal systems and the minerals form homogeneous grains or crystals. Crystalline minerals develop best by slow precipitation in solutions contained in open spaces; under such conditions free crystals may form in silicate melts or magmas and in aqueous solutions. Crusts of minerals may develop where crystals adhere to walls of water filled fissures and this is a common feature in mineral veins. The first impulse to crystallization may be given by adsorption and supersaturation along the walls. Once started the larger crystals become further enlarged because smaller crystals dissolve more rapidly than large ones and the liquid remains supersaturated with reference to the larger growths.[1]

When crystallization is progressing from a great number of points in the solution, a granular texture is developed by the mutual interference of the crystals. In mineral deposits the resulting textures are usually rather coarse; only very rarely do we find fine-grained aggregates of order of magnitude of the dense groundmass of igneous porphyritic rocks. In mineral deposits a considerable variety of textures result by replacement in solid rock; such textures are in general analogous to those found in metamorphic rocks.

Colloids.[2]—A number of minerals are formed both as crystalloids and as colloids. The usage requires a different name for the two forms. The opinion is gaining ground that the colloid state is not a separate kind of matter but simply crystalloids in a state of dispersion ranging from comparatively coarse suspension down to almost molecular subdivisions. A colloid mixture is thus a two-phase, heterogeneous system in which the solid, divided into small separate volumes is known as the disperse phase and the liquid as the dispersion medium. The disperse phase may separate from the mixture in a gelatinous or flocculent form, still retaining some of the solvent. In one class of these mixtures, called colloidal solutions, a gelatinous mass is obtained by cooling or evaporation; this is termed a "gel." Silica in aqueous solution, separating as opal is one of the more common of these gels occurring in nature. Gelatinized colloids are permeable to salts, the rate of diffusion being almost the same as in water, but they diffuse themselves through other colloids or through porous walls with the utmost difficulty. The colloidal solutions are not coagulated by salts.

Another class of colloidal mixtures, known as "colloidal suspensions" is more abundantly represented in nature; it is believed that in these the suspended particles are of much larger

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[1] W. Ostwald, The scientific foundations of analytical chemistry, 1900, p. 22.

[2] Arthur A. Noyes, The preparation and properties of colloidal mixtures. *Jour.* Am. Chem. Soc., vol. 27, 1905.

W. Ostwald, Handbook of colloid chemistry, 1915.

size than in the colloidal solutions. These colloidal suspensions are not viscous and do not gelatinize but coagulate readily when an electrolyte is added. They result from reactions between two chemical compounds in the absence of electrolytes. When, for instance, hydrogen sulphide is added to a solution of arsenious oxide, a turbid yellow liquid results which is a colloidal suspension of arsenic sulphide. An important fact is that a small quantity of a gelatinizing colloid prevents the coagulation of colloidal suspensions by electrolytes. Thus a small amount of gelatine may keep silver chloride indefinitely in a state of colloidal suspension.

The importance of these facts as related to reactions in natural solutions is obvious. Many waters are rich in colloidal silica and its influence may prevent a precipitation which otherwise would take place. Another effect is the retention of silica jelly within the walls of fissures while gases and electrolytes pass through them.

The existence of colloid minerals was recognized by the German mineralogist Breithaupt, and Berzelius already knew of the properties of many colloids. Grahams studies about 1864 laid the real foundation for colloidal chemistry, a branch which, however, has only recently begun to attract the attention it merits. Colloidal minerals form in great abundance at ordinary temperatures, near the surface, within the oxidized zone. Some colloids like opal may be deposited at higher temperature of 100° C. to 200° C. perhaps even higher, but this is exceptional. The solidified colloids have a great tendency to acquire crystallinity and in time—rapidly with heat—become transformed into fibrous or cryptocrystalline aggregates. Such crystallized colloids may be called metacolloids.[1] Almost all metals, oxides, hydroxides and sulphides may yield colloidal solutions or suspensions. Among minerals of colloidal origin common in nature may be mentioned opal, limonite, psilomelane, calcium phosphate, arsenic and perhaps some secondary varieties of chalcocite.[2]

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[1] E. Wherry, *Jour.* Washington Acad. Sci., vol. 4, 1914, p. 112.

[2] F. Cornu, Many papers in. *Zeit.* f. Chem. u. Ind. d. Colloide, vol. 4, 1909.

E. Wherry, *Op. cit.*

Austin F. Rogers, A review of the amorphous minerals, *Jour. Geology*, vol. 25, 1917, pp. 515–541.

CHAPTER III

THE FLOW OF UNDERGROUND WATER

General Statement.[1]—Much of the rain water which descends upon the land runs off in surface drainage, and a smaller part evaporates, but a certain quantity sinks down into the soil and into the porous and fractured rocks. This part of the precipitation adds to the *ground water* and, if there is a sufficiency of rainfall, it saturates the material at a certain varying depth below the surface. This upper limit of the saturated zone indicated by the depth at which water stands in wells or shafts is called the water level, ground-water level, watertable, or hydrostatic level. It is as a matter of fact a warped surface which feebly reflects the topographic features (Fig. 1). The water may penetrate to considerable depths, particularly along fissures; gravity and heat often establish a circulation of the ground water

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[1] C. R. Van Hise, A treatise on metamorphism, *Mon.*, 47, U. S. Geol. Survey, 1894, pp. 123–191, and 657–670.

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J. F. Kemp, Igneous rocks and circulating waters, etc., *Idem.*, vol. 33, pp. 707–711.

J. F. Kemp, The problem of metalliferous veins, *Econ. Geol.*, vol. 1, 1906, pp. 207–232.

J. W. Finch, The circulation of underground aqueous solutions and the deposition of lode ores, *Proc.*, Colo. Sci. Soc., vol. 7, 1904, pp. 193–252.

M. L. Fuller, Total amount of free water in the earth's crust, *Water-Supply Paper* 160, U. S. Geol. Survey, 1906, pp. 59–72.

T. A. Rickard, *Eng. and Min. Jour.*, March 14, 1903; *Min. and Sci. Press*, June 27, 1908.

so that it may again reach the surface after a long underground journey. In all its aspects the ground water performs important geologic work by solution and precipitation, which may result in the concentration of useful minerals.

There are other waters which may be found in the crust: The sea water may at times invade the rocks, or may have been occluded in old sediments. There is a small class of hot waters, ascending from great depths which are believed by many geologists to be of magmatic origin, that is given off by cooling magmas. However important such waters may be in the formation of certain kinds of mineral deposits, they are insignificant in quantity compared to the great mass of water of atmospheric origin which is contained in the rocks.

Pores and Openings in Rocks.—All rocks are porous and are capable of absorbing water. By porosity is understood the percentage of pore space referred to the total volume of the rock. The ratio of absorption is the ratio between the weight of the water absorbed and the weight of the rock tested. When the pores are completely filled the rock is said to be saturated, but a saturated rock after being drained always retains a certain amount of water which adheres to the walls of the pores. The pore space varies from a fraction of 1 per cent. to 40 per cent. In fresh granites and similar compact rocks the porosity is from 0.2 to 0.5, in limestones from 0.53 to 13.36, in sandstones from 5 to 28 per cent.

Under the assumption that a sandstone consists of spherical grains packed in the most compact arrangement possible, the space between the spheres would amount to 25.95 per cent. Other things being equal the porosity increases with the size of the grains.

In loose sand and gravel the porosity is highest, ranging from 32 to 40. The absorbed water may be called "free water" in contrast to that existing in chemical combination in the minerals of the rock. All of the free water is not "available," for instance in wells, because some rocks, like clays, have the peculiarity of holding in their pores great quantities of water which is released only at an extremely slow rate.

On the basis of size, openings in rocks may be divided into (1) openings which are larger than those of capillary size, or supercapillary openings; (2) capillary openings; and (3) sub-capillary openings. For water, openings larger than capillary openings may be considered as circular tubes which exceed 0.508 mm. in diameter, or sheet openings, such as those furnished by faults, joints, etc., whose width exceeds 0.254 mm. Capillary tubes or sheet spaces are those smaller than the dimensions indicated but larger than the openings in which the molecular attraction of the solid material extends across the space, and to such openings the laws of capillary flow apply. In sub-capillary openings the attraction of the molecules extends from wall to wall, and this class includes tubes smaller than 0.0002 mm. in diameter and sheet openings smaller than 0.0001 mm. in width. According to G. Bakker[1] the spheres of molecular attraction are only six to seven times the molecular diameter and consequently, if this be true, capillary movement can take place in tubes very much smaller than 0.0002 mm.

The flowage of water through super-capillary openings nearly follows the ordinary laws of hydrostatics, but is subject to a certain retardation on account of friction. The super-capillary openings include the greater number of faults, joints, partings, and the openings in coarser sediments. In capillary openings the movement is very slow indeed, so that many rocks in which they occur, as shales and clays, are spoken of as impermeable. In sub-capillary openings the water is held firmly as a film glued to the walls by adhesion; there is no free water and the flow is practically nil.

The capillary forces play a considerable part in the movement of underground waters, but, in general, they cannot produce a continuous flow, and they are of secondary importance in comparison with the hydrostatic pressure. At higher temperatures the capillary action decreases and becomes zero at the critical points.

Openings in rocks do not persist indefinitely in depth, though very hard rocks like granite will sustain far greater loads than at the surface. The experiments of F. D. Adams[2] indicate that cavities may exist in granite to a depth of at least 11 miles, or 17,600 meters. Most rocks tend to become plastic at far lesser depths and will then become deformed and flow without fracture. Under such conditions an active circulation of water is difficult.

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[1] *Zeitschr. f. phys. chem.*, vol. 80, 1912, p. 129. See also J. Johnston and L. H. Adams, *Jour. Geology*, vol. 22, 1914, p. 13.

[2] *Jour. Geology*, vol. 20, 1912, pp. 97–118.

Water in Sands and Gravels.—We may first consider the simpler case, of loose material such as sands and gravels which are so abundant in the uppermost part of the crust, and in comparison with which the underlying compact rocks may be regarded as impermeable. Under the influence of gravity the water descends until a depth is reached where the material is saturated, this being the water level. In the valleys the water level will lie close to the ground while it rises slightly under the ridges so that it may be necessary to sink deeply for wells with permanent water. The ground water is not stationary but moves slowly from the higher ground toward the valleys; and underneath the valleys its perceptible movement continues down stream until ultimately, with lack of grade, for instance, where a river valley opens toward the sea, the movement will become slow and almost imperceptible. That such a movement actually takes place has been proved by the use of fluorescein and other indicators in wells and bore holes.

Water in Rocks of Uniform Texture.—In uniform rocks, like granite, water is not only contained in the pores but moves more easily on the ever present seams and joints. Considerable water may be stored, though the quantity per unit volume of rock will be much smaller than in sands and gravels. The water table is here also a curved surface which follows approximately the topographic relief but is less accentuated. The elevation of the water table or water level may fluctuate considerably dependent upon seasonal rainfall.

In his paper, cited above, J. W. Finch distinguishes the space above the water level as the *gathering zone* or *zone of percolation*, in which water and air are both present, and in which the water is conducted to the *saturated belt* (Fig. 1). Even in very arid regions there is usually a deep zone of saturation, though the quantity of water stored may be small. It would also be possible, however, to have no belt of saturation and the water would then simply percolate feebly downward until the quantity is diminished to zero.

The *zone of discharge* "embraces that part of the belt of saturation which has a means of horizontal escape" by continuous gravitative flow.[1] The movement of the water is usually more rapid in the upper part of this zone than in the lower part, where, in spite of greater pressure, the obstructions and the increasing

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[1] *Proc.*, Colorado Sci. Soc., vol. 7, 1904, p. 202.

compactness of the rock retard the flow. The air or gas filling the pore space must also be driven out before the water can enter, and the evaporation of water in the underground atmosphere may, under some conditions, also become an important factor in reducing the supply of water. Still another portion of the water enters into chemical combination by hydration.

The *static zone* is the third and deepest of the divisions proposed by Finch. It extends below the level of the lowest point of discharge and the water in it is stagnant or moves extremely slowly. It depends upon the zone of discharge for its water, as it is simply the bottom part, with gradually diminishing quantity of water, of a belt of saturation of which the zone of discharge is the upper and flowing part. The lower limit of the third zone, where the quantity of water becomes exceedingly small, is not entirely a matter of speculation, for many definite data are supplied by mining operations and in many places it is not more than 1,500 feet below the surface. Large quantities of water may be stored in the third zone, as, for instance, in the deep artesian basins where impermeable beds prevent escape. A certain amount of "rock moisture" undoubtedly persists to great depths.

Water in Sedimentary Rocks.—In a series of sedimentary beds it is common to find impermeable rocks like clay and shale alternate with more porous beds like sandstone and limestone. Under such conditions the distribution of the de-

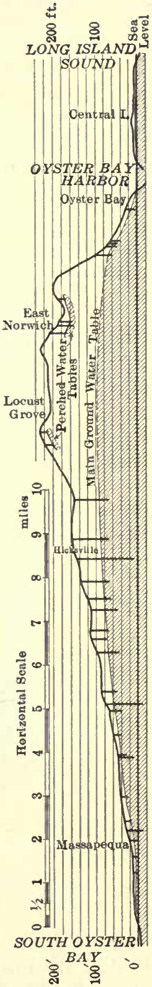

FIG. 1.—Section across Long Island, showing actual position of ground-water table. Upper part of section shows zone of gathering with local ground-water tables due to presence of impermeable strata. Zone of discharge between ground-water table and sea level. Static zone below sea level. *From Prof. Paper 44, U. S. Geol. Survey.*

scending surface waters is likely to be irregular. Near the surface there may be a local water level but below this beds heavily charged with water may alternate with almost dry strata. Each bed may in a way be considered as a unit and if it outcrops it has its own zone of gathering, zone of discharge and its static zone. The Cretaceous Dakota sandstone presents an excellent example of a porous stratum in which a large amount of water can be stored. Throughout the Great Plains this is a veritable reservoir of water, which can be tapped by artesian wells as far as 300 miles from its outcrop and at depths of a few hundred to 3,000 feet (Fig. 2). But this stratum at present simply contains a stagnant body of water, and, as in most other artesian basins, the quantity is not inexhaustible. This very case proves how impervious the adjacent sedimentary beds are, for neither upward nor downward is an avenue of escape afforded in spite of the strong pressure. Should profound fissuring take place in the Great Plains a natural avenue of escape would, of course, be opened and a deep circulation established. Kemp and Fuller have both brought out the fact that the deep sedimentary beds are often remarkably dry. The well 4,262 feet deep at Wheeling, West Virginia, was in absolutely dry rocks for the lower 1,500 feet. Wells sunk at

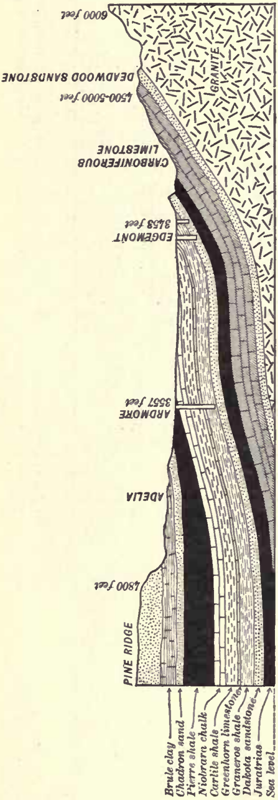

FIG. 2.—Section from the Black Hills, South Dakota, to Pine Ridge, across Fall River County, showing relation of the water-bearing Dakota and associated sandstones, looking west. Vertical scale exaggerated. *After N. H. Darton, U. S. Geol. Survey.*

Northampton, Massachusetts, and at New Haven, Connecticut, to depths of 4,000 feet have failed to obtain water. A number of other instances are mentioned, and in many cases the dry part consists of sandstones or other porous rocks.

Some time ago it was suggested by A. C. Lane[1] that part of the salt water in deeply buried beds is fossil sea water or "connate" water occluded in the sediments at the time of deposition. There can be little doubt that dryness as well as salinity increase with depth.[2]

Influence of Fractures.—The simple conditions outlined above are seriously disturbed where extensive fracturing has taken place and paths have been laid out on which the water may move under approximately normal hydrostatic conditions. There may be a comparatively slow descent of the water along devious joints and fractures and a rapid rise under hydrostatic head where the descending water reaches the open paths of important faults and fissures. Friction during the descent undoubtedly seriously diminishes the theoretical head, but the evidence is perfectly clear that in regions of dynamic disturbance, such as the Alps and the Rocky Mountains, strong ascending springs may result from these conditions.

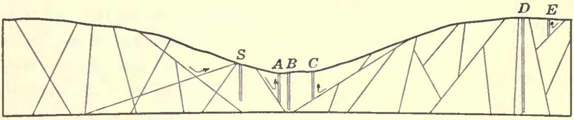

Fig. 3.—Section illustrating flow of water in jointed crystalline rocks. *A*, *C*, flowing wells fed by joints; *B*, intermediate well of greater depth, with no water; *D*, deep well not encountering joints; *E*, pump well adjacent to *D*, obtaining water at shallow depth; *S*, dry hole adjacent to spring. *After M. L. Fuller, U. S. Geol. Survey.*

At the point of issue such springs may be warm and their temperature, in regions where no recent igneous action has taken place, may be a good indication of the depth attained by the water. Such springs seldom have a temperature higher than 65° C., and the composition of their salts corresponds to the character of the beds traversed. On the supposition, believed to be well founded, that only a moderate loss in heat takes place

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[1] *Bull.* Geol. Soc. Am., vol. 19, 1908, p. 502.

[2] See Chapter VI.

during the ascent, a water of the temperature named would be derived from a depth of about 5,500 feet. Large regions of the earth, such as the Scandinavian peninsula, contain no warm springs, and the eastern part of the American Continent yields very few of them. Fuller says:

> The results of drilling in sedimentary and crystalline rocks, as well as the studies of deep mines, show that in all probability water does not commonly exist in the rocks under great pressure, although such may be exerted in an occasional crevice. It is not believed that hydrostatic waters exist, except possibly in rare instances, at depths of over 10,000 feet, and that in reality the estimate of a depth of 6,520 meters, or 20,000 feet, as the limit of the zone of open cavities is closely approximate to the truth. . . . . . If waters were freely circulating at great depths, within the zone of fracture, hot springs would certainly be more common along the numerous faults of the Piedmont, Appalachian, and similar regions.[1]

Van Hise suggests that the decreased density and viscosity of water at higher temperatures may lessen the head necessary for ascending springs, but it may be doubted whether these factors would ever offset the great friction encountered during the downward passage. Faulting and mountain-building processes develop heat and this disturbance of the conditions of temperature may result in convection currents and an increased circulation of the water stored in the rocks.

Influence of Volcanism.—When magmas are intruded into the zone of fracture the conditions become more complicated. It is thought by some that atmospheric waters are able to descend into the deep regions and become absorbed by the magmas, but this view appears improbable. Before its irruption into the zone of fracture the magma is assuredly far beyond the reach of any waters percolating from the surface. Daubrée's well-known experiment has often been cited, as showing how water may pass through a disc of sandstone against a certain counter pressure of steam. Recent critical examination[2] has shown the fallacies involved in the experiment, and indicate that "the probabilities are all against the notion that appreciable amounts of meteoric waters can ever penetrate into deep-seated and highly heated rock masses."

[1] *Water-Supply Paper* 160, U. S. Geol. Survey, 1906, p. 64.

[2] John Johnston and L. D. Adams, Observations on the Daubrée experiment, etc., *Jour. Geology,* vol. 22, 1914, pp. 1–15.

The presence of a heated body in the zone of fracture would undoubtedly quicken the circulation of water by inducing strong convection currents and expelling the stored water from its reservoir. Whether this action is sufficient to account for the remarkable number and volume of hot springs rising in volcanic regions may well be doubted, and it is thought that the magma itself gives up most of its constitutional water, partly when moving up to higher levels, partly when crystallizing to solid rocks.

Conclusions.—In conclusion it is believed that water in quantities sufficient to supply an ascending circulation can only exceptionally attain a depth of 10,000 feet and that, except in regions of great dynamic movements, the active circulation is confined to the uppermost few thousand feet. More commonly the depth of active circulation is measured by the level of surface discharge and the water below that level is practically stagnant; the lower limit of the body of stagnant water then forms an irregular surface descending to greater depths along the fractures and rising higher in the intervening blocks of solid ground.

Examples of Movement of Water.—According to Fuller, water supplies in wells in crystalline rocks are usually found within 200 or 300 feet of the surface and it is ordinarily useless to go below a depth of 500 feet (Fig. 3). The occurrence of porous strata which are capable of holding immense quantities of water but in which none whatever is actually found is, according to Fuller, a common experience of drillers in this country, even where the upper strata contain a well-defined water table.

Investigations of joints in the crystalline rocks of Connecticut have shown, according to Fuller, that the water occurs largely in the vertical joints, which have an average spacing of 3 to 7 feet at the surface. In depth these joints diminish rapidly or close up and it is therefore not advisable to go below 250 feet in search of water. It is estimated by E. E. Ellis that the water present in the upper 2,000 feet of the crystalline rocks is only 16 per cent. of their capacity, or 0.000007 of the rock volume.

The evidence from many mining regions is of considerable importance. In the Sierra Nevada of California deep canyons are separated by broad-backed ridges capped with Tertiary gravels and andesitic tuffs. The abundant precipitation percolates into the porous tuffs and gathers in the gravel basins, from the lower parts of which large quantities of cold springs issue. This upper zone of gathering and discharge may be 1,500 feet deep and may lie the same distance above the bottom of the canyons. In spite of the depth of the percolating zone, the waters are potable, pure, and cold. A part of the water sinks into the underlying bedrock of slate or granite, but the quantity is far less than in the more porous Tertiary strata and it finds its lowest level of discharge along the beds of the rivers. For the Sierra as a whole the Great Valley of California forms the ultimate level of discharge. In the whole western part of the range there are no thermal springs and very few strong ascending springs, in spite of the prevalent fissility and jointing in the rocks. Hot springs are encountered only along the eastern slope of the range, a region which in the late Tertiary and Quaternary was the scene of great dislocations and volcanic activity. In the gold-quartz veins contained in the old rocks of the western slope much water is found in fissures to a depth of about 800 or 1,000 feet. Below this little water is met and many stopes and drifts are entirely dry, and this applies both to mines high up on the slopes, as at Nevada City and Grass Valley, and to the Mother Lode mines of the foot-hill region.

Cripple Creek, Colorado, is another interesting example. Here we have a granitic plateau at an elevation of 9,000 feet above the sea; this plateau contains a volcanic plug about 2 miles in diameter which is largely filled with porous breccias and tuffs. The water fills the volcanic rocks as in a sponge inserted in a cup and the mining operations to a depth of 1,500 feet have tapped heavy flows. But even in this water-logged mass there are solid intrusive bodies, for instance at the Vindicator mine, at a depth of 1,000 feet, which are so dry that water must be sent down for drilling. The data thus far available have led Ransome to the conclusion that even at Cripple Creek the water is slowly diminishing in quantity at increasing depth. The big drainage tunnel now under way, which will tap the veins at a depth of 800 feet below the present lowest tunnels, will afford more information on this subject. The granite which surrounds this water-soaked plug contains very little water and at most places is practically dry, in spite of the great hydrostatic pressure. The ultimate level of possible discharge would be in the valley of the Arkansas, 2,500 feet lower and many miles distant, but it may be gravely doubted whether any water from the Cripple Creek mines ever finds its way through the granite mass to this level.[1] Van Hise, after stating (Metamorphism, p. 1065) that during a certain time the Portland mine, at Cripple Creek, yielded water to the amount of between 300 and 900 gallons per minute, asks whether better evidence could be required for proving the existence of an extremely active circulation. The answer to this is that the water was simply stagnant, stored water filling an underground reservoir.

In the copper mines of Butte, Montana, where the granitic rocks are greatly faulted by movements of late date, much water was encountered, extending in places down to 2,400 feet, or the bottom of the mines. No ascending springs are found at the surface, nor any hot springs, although a high range adjoins the mines on the east and conditions seem to be favorable for deep circulation. The water is probably almost stagnant, and Weed mentions the existence of large bodies of dry rock.[2] One such body on the 1,600-foot level, 1,200 feet in width, is absolutely dry.

Leadville, Colorado, is another place where the faulting is extensive and of comparatively recent date. At 1,500 feet, the greatest depth attained, there is still much water, mainly along the faults.

At Rossland, British Columbia, according to Bernard McDonald,[3] the mine waters increase greatly during the spring months. The water level is at 40 feet and the quantity increases to a depth of 200 to 350 feet. Below 350 feet a decrease begins, slowly at first but soon more rapid, until at 900 feet there is only a slight seepage and below 1,000 feet the mine is dry.[4] Weed states that in the copper mine at Ely, Vermont, an incline shaft was carried down for a length of 3,600 feet, attaining a vertical depth of 1,700 feet. There is no water here below a vertical depth of 600 feet. At Przibram, Bohemia, the workings are dry and dusty at a depth of 3,000 feet. In Cornwall, and in New Foundland mines have been worked underneath the sea, and sometimes close to the sea bottom, without irruptions of salt water.[5]

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[1] Lindgren and Ransome, *Prof. Paper* 54, U. S. Geol. Survey, 1906, pp. 233–251.

Finch, J. W., *op. cit.*, p. 204.

[2] M. L. Fuller, *Water-Supply Paper* 160, U. S. Geol. Survey, 1906, p. 65.

[3] T. A. Rickard, *Min. and Sci. Press*, June 27, 1908.

[4] M. L. Fuller, *op. cit.*, p. 65.

[5] For other examples see J. F. Kemp, The ground waters, *Trans.*, Am. Inst. Min. Eng., vol. 45, 1914, pp. 3–25.

One of the most convincing examples is that furnished by the deep copper mines of Michigan and fully set forth by A. C. Lane.[1] He shows that the surface waters are of the normal, potable type and that they descend in diminishing quantities only to a depth of about 1,000 or 1,500 feet below the surface. Below this depth moisture is scant, but where it appears it consists of drippings of strong calcium chloride brine which cannot in any way be explained as being derived from the surface water. Many levels are absolutely dry and water must be sent down for drilling. This case is particularly convincing, for we have here many features in favor of a strong circulation: Moist climate, inclined position of beds, and great permeability.

No certain figure can be assigned to the depth of the ground water; it may be shallow or the water may descend on strong fractures for several thousand feet. At any rate the quantity is limited, and the water is largely stagnant and is much more likely to decrease than to increase at depths below 1,000 feet.

Depth of Water Level.—In moist climates the water level is usually found within 50 feet of the surface, but in regions with less rainfall there is great diversity in the location of this upper limit of the zone of saturation. In the more arid regions the water is often met 300 or 400 feet below the surface. In the valley of Hachita, New Mexico, no water is found in the sands and gravels until a depth of 500 feet is reached; at the Abe Lincoln gold-quartz mine, New Mexico, a little water began to come in 1,300 feet below the surface. In the rich deposits of Tintic, Utah, the water level in limestone lies 1,700 to 2,000 feet below the surface, but in mines in andesite and porphyry in the same district water may be found at much less depth.

When water is being drained or pumped from a mine the water level is artificially depressed, in the form of a flat funnel. The pump in this case does not merely drain the bottom level, but receives water from higher levels farther away from the shaft. It is important to note this, for the water thus obtained from the bottom of a wet mine may not have the same composition as that originally belonging to this level.

Total Amount of Free Water in Earth's Crust.—Several estimates have been made of the total amount of uncombined water contained in the upper crust. The older estimates by Delesse,

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[1] Mine waters, *Trans.*, Lake Superior Min. Inst., vol. 12, 1908, pp. 154–163.

Dana, and Slichter were very high. Chamberlin and Salisbury[1] believed that the water in the earth would be equivalent to a layer 800 feet deep over its entire surface. Van Hise[2] reduced the estimate materially and concluded that it would be equivalent to a sheet of water 226 feet (69 meters) thick over the continental areas.

Fuller[3] estimates, after a careful study of the problem, that the total water would be equivalent to a uniform sheet 96 feet thick over the entire surface of the earth. This estimate is probably more nearly correct than any of the others.

[1] Geology, 1904, vol. 1, pp. 206–207.

[2] A treatise on metamorphism, *Mon.* 47, U. S. Geol. Survey, 1904, pp. 128–129, 570–571.

[3] *Op. cit.*, p. 72.

CHAPTER IV

THE COMPOSITION OF UNDERGROUND WATERS

INTRODUCTION

Water is continually evaporated from sea and land. From the gathered clouds it is precipitated as pure rain water, which, by the aid of absorbed oxygen and carbon dioxide immediately begins the attack on disintegrating rocks. The rivers finally carry suspended particles and dissolved salts to the sea. Considered more closely, the rivers are the products of the weak solutions of the immediate run-off and the stronger ground water solutions from the zone of discharge (Chapter III) which have been in longer contact with the rocks and leached them more thoroughly. This inconspicuous process of decomposition of rocks and solution of resulting salts is one of scarcely realized geologic importance. F. W. Clarke[1] states that the Mississippi annually carries to the sea about 108 metric tons of salts from each square mile of territory drained; the Colorado abstracts about 51 tons from the same unit area.

A smaller part of the ground water sinks to enrich the static zone of stagnant waters, and ultimately becomes highly charged with salts. A still smaller part of the water is permanently withdrawn by entering hydrated compounds like kaolin.

To complete the picture we must not overlook the ascending hot solutions which come from great depths and which in part at least are derived from rising magmas.

As most mineral deposits have been formed by aqueous solutions the composition of the waters of rivers, lakes and seas becomes a study of importance. Even more important is the composition of the underground waters of wells and springs. In considering them from a chemical standpoint it will be best

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[1] F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 113.

R. B. Dole and H. Stabler, *Water-Supply Paper* 234, U. S. Geol. Survey, 1909, p. 78.

to attempt no artificial distinction between thermal or cold mineral or non-mineral waters.

The substances dissolved in the ground water depend upon the formations which it traverses. At the immediate surface organic life may influence the composition. In general, each formation yields its characteristic salts to the precolating waters. Each natural water is a chemical system of balanced constituents of more or less dissociated electrolytes, of colloids, and of gases.

CALCIUM CARBONATE WATERS IN IGNEOUS ROCKS

Igneous rocks, of deep-seated origin, as well as crystalline schists contain only small amounts of soluble salts. The surface waters penetrating them are charged with more or less carbon dioxide, which, at ordinary temperatures, gradually decomposes the silicates, particularly the pyroxene, amphibole, biotite, and the calcium feldspars; the alkali feldspars are more slowly attacked. As a result the springs in such terranes will have a low salinity, rarely above 1,000 parts per million, and will contain principally calcium carbonate, with more or less of the corresponding magnesium salt; a smaller amount of sodium carbonate and much less of potassium carbonate are present. There will be little of the chlorine and sulphuric acid radicles. The silica is relatively high. Such calcium carbonate waters are characteristic not only of superficial springs, but also of the deeper circulation in crystalline terranes; in the latter case the waters may be warm, though usually they are cold. The spring-fed rivers in such terranes have a similar composition.

Where magnesian rocks like basalt and serpentine abound, the underground waters are richer in magnesia than usual, and this substance may even equal the calcium. Waters of this calcium carbonate type are common and, when encountered as ascending springs or elsewhere, justify the presumption of surface origin.[1]

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[1] Some years ago, in a report on the gold-quartz veins of Nevada City and Grass Valley, Cal. (*Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, p. 121), I presented an analysis of an ascending spring found in the Federal Loan mine which carried some arsenic and hydrogen sulphide. At that time I held the opinion that this water might possibly have had some connection with the genesis of the vein, but it is now apparent that it is simply water of the general surface circulation which happened to find its way up on the vein and which dissolved certain constituents from it. The analysis is quoted on page 44.

It often happens that hot springs which are not characterized by an abundance of calcium carbonate are accompanied by numerous other springs of somewhat lower temperature. A comparison of analyses will usually show that in proportion to the lowering of the temperature the quantity of calcium carbonate increases; this indicates a cooling admixture of surface waters bearing calcium.

COMPOSITION OF SALTS AND TOTAL SALINITY OF SURFACE WATERS IN CRYSTALLINE ROCKS

	A	*B*	*C*	*D*
CO_3	31.91	47.14	57.80	38.46
SO_4	9.07	6.67	3.10	15.35
Cl	4.03	4.18	1.30	2.81
S			0.50	
Ca	14.53	22.67	13.90	13.24
Mg	2.93	6.17	2.30	4.33
Mn			0.10	
Na	10.80	5.32	5.50	12.86
K	2.72		0.40	3.76
$(Al,Fe)_2O_3$	0.51		1.70	
SiO_2	23.50	7.85	13.40	9.19
	100.00	100.00	100.00	100.00
Salinity, parts per million	37	280	245	282

NOTES RELATING TO ABOVE ANALYSES

A. Cache la Poudre River, Colorado, above North Fork. In schist and granite. Analysis by W. P. Headden. See F. W. Clarke, Geochemistry, Bull. 616, 1916, p. 65.

B. Aztec Spring, 4 miles east of Santa Fe, New Mexico. In schist and granite. Cold. Analysis by F. W. Clarke, Geochemistry, p. 64.

C. Cold spring in Federal Loan mine, in granite and schist, Nevada City, California. Approximate analysis by W. F. Hillebrand; contains a little manganese and trace of lead. Deposits calcium carbonate, limonite, and some arsenic. Sulphur probably due to reduction from small amount of H_2S after bottling. *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, p. 121.

D. Cold water from 500-foot level, Geyser mine, Silver Cliff, Colorado. Analysis by W. F. Hillebrand. In "The mines of Custer County, Colorado," by S. F. Emmons. *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, p. 461. Free and, semi-combined CO, 38.8 parts per million.

Waters of the kind described above are generally poor in the rarer metals. A little arsenic is found in some cases and traces of barium, strontium, lithium, boron and phosphorus are sometimes recorded. Where they traverse mineral deposits, metals contained in the deposits will of course be dissolved, as in the water from the Federal Loan mine, Nevada City, California. The springs may, under favorable conditions, form crusts of calcium carbonate and hydroxide of iron, but as a rule their powers of solution and deposition are weak. Where the rocks contain much pyrite, as often is the case in mining districts, the sulphates, especially calcium sulphate, rapidly increase in the waters.

CALCIUM CARBONATE WATERS IN SEDIMENTARY ROCKS

Waters of the type described above are not confined to igneous rocks. They are often found in circulation in glacial drift and also in sedimentary rocks—sandstones, limestones, and dolomites. Such waters sometimes contain hydrogen sulphide and carbon dioxide. The derivation of the latter is not always easily explained. In some cases the gas may emanate from a deep-seated magma, but more commonly it is formed by decomposition of carbonates. An example of such water is furnished by the cold Cresson Spring in Pennsylvania, which issues from a shale member between sandstones in a 3,000-foot series of Coal Measures, containing practically no limestone. This water is pure, its salinity being only 442 parts per million, and of this 272 may be calculated as calcium carbonate, 76 as sulphates of sodium, magnesium, and calcium, and 11 as sodium chloride. According to a careful analysis by Genth this water contains traces of nickel, cobalt, iron, manganese, copper, strontium, barium, and fluorine, 0.17 part per million of the last-named element being present. Several analyses of similar well-known waters are quoted on page 46.

COMPOSITION OF SALTS AND TOTAL SALINITY OF CALCIUM CARBONATE WATERS IN SEDIMENTARY ROCKS

	A	*B*	*C*
CO_3	40.02	41.47	[1]48.64
SO_4	21.73	3.93	6.30
Cl	0.64	1.27	5.40
PO_4	....	0.03	....
NO_3	....	0.23	....
Ca	23.35	23.54	16.56
Mg	5.82	2.56	7.64
Na	1.81	2.38	10.36
K	2.04	0.80	....
NH_4	....	0.03	....
Mn	....	0.17	....
Fe_2O_3	....	0.10	0.40
Al_2O_3	0.58		0.20
SiO_2	4.01	22.85	4.50
BO_2	....	0.64	....
	100.00	100.00	100.00
Salinity, parts per million	563	199	222

[1] HCO_3.

NOTES RELATING TO ABOVE ANALYSES

A. Virginia Hot Springs, Virginia. Analysis by F. W. Clarke. "Geochemistry," p. 193. Temperature tepid. Issues from Paleozoic sediments. See also *Bull.* 32, U. S. Geol. Survey, 1886, p. 61.

B. Hot Springs, Arkansas. Spring No. 16. Temperature 62° C. Issues from sharply compressed folds of Silurian sandstone and shale. CO_2 from bicarbonates 28.34 cc. per liter; nitrogen 8.39 cc. per liter; oxygen 2.49 cc. per liter; H_2S none. Arsenic none; trace iodine and bromine. Analysis by J. K. Haywood. The Hot Springs of Arkansas, *Senate Doc.* 282, Fifty-seventh Congress, First Session, 1902, p. 94. Recalculated by F. W. Clarke, "Geochemistry," 1916, p. 195.

C. Cold water from well of Missouri Lead and Zinc Company, Joplin, Missouri. Depth 1,387 feet. In Paleozoic limestone. Analysis by Cleveland and Millar Laboratory. *Water-Supply Paper* 195, U. S. Geol. Survey, 1907, p. 137. Recalculated.

These waters frequently form ascending springs. Bartlett Springs, Lake County, California, the water of which is extensively used in that State, probably belong to this class. The water contains 782 parts of salts per million, of which 493 may be calculated as calcium carbonate. It is rich in free carbon dioxide and is low in chlorine, sulphuric acid radicle, sodium, and potassium, but contains some iron, probably as carbonate, a little barium, phosphorus, and about 63 parts of silica per million.[1]

Carbonate waters are undoubtedly active in solution and deposition in the upper part of the crust, and especially in the formation of concentrations from weathering rocks. They may deposit calcareous sinters and effect concentrations of iron and manganese. Some lead and zinc deposits in limestone may also be genetically connected with them; their power of solution and concentration of rarer metals appears to be weak, unless they contain carbon dioxide and hydrogen sulphide. Such waters in Kansas, Missouri, and Kentucky have been found to contain zinc and probably also lead and copper.

The salts are surely obtained from the rocks traversed.

CHLORIDE WATERS IN SEDIMENTARY ROCKS

Infiltration from Present Oceans.—Wells and springs along the sea coasts usually contain a higher percentage of sodium chloride than farther inland; this may be caused either by infiltration of sea water into sediments or porous igneous rocks, or by winds carrying finely divided salt from the spray of the waves.

Solution of Saline Deposits.—Many past geologic periods included epochs of desiccation and desert climate when salt was precipitated from evaporating waters of closed basins. Surface waters encountering such sedimentary deposits easily dissolve the sodium chloride, and wells and springs rich in this salt are characteristic of many regions. Besides sodium these waters contain calcium and magnesium, and they are often rich in calcium chloride. They are poor in silica and potassium and rarely contain much calcium which can be combined with carbon dioxide. The presence of bromine is almost characteristic; traces of iodine and boron are often found. Barium and strontium are almost always present, the former sometimes in considerable amount. Free carbon dioxide and hydrogen

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[1] Winslow Anderson, Mineral springs, etc., of California, 1892, p. 94. G. A. Waring, Springs of California, *Water-Supply Paper* 338, U. S. Geol. Survey, 1915.

sulphide are sometimes found, the latter especially where there is an abundance of calcium sulphate. Waters of this general type are characteristic of certain Paleozoic beds in the eastern United States, as, for instance, the Silurian of New York and Michigan and certain parts of the Carboniferous in Michigan. In the western States the "Red Beds," generally of Permian or Triassic age, are sometimes rich in salt and gypsum, and this combination appears in the waters of these terranes.

There are many similar springs and wells in Pennsylvania, and in fact all through the interior Paleozoic basin, from Arkansas to Canada. The Saratoga Springs of New York, issuing from Silurian limestones, probably belong to this class. Their temperature is about 50° F.; the total solids amount to about 11,000 parts per million, of which the larger part is sodium chloride. Barium is conspicuously present, in some analyses to a maximum of about 34 parts per million, likewise bromine at about 1.20 parts per million. Small amounts of silica, iron, and lithium, and traces of boron, iodine, and fluorine are recorded. The origin of the CO_2 so abundant at Saratoga Springs is uncertain. J. F. Kemp believes it to be of magmatic derivation. Examples of such waters are given in the table of analyses on page 50.

Certain of these waters are abnormally rich in calcium chloride, that most easily soluble salt which remains as the last liquid residue in evaporating brines. Several instances of such waters have been interpreted as residual or connate brines, remaining in early isolated Paleozoic basins.[1]

In the lower peninsula of Michigan brines are obtained from deep wells in the Carboniferous and Silurian. One of the springs in this region contains 12,000 parts per million in total solids, with 6,000 calculated as NaCl, 1,600 as $MgCl_2$, and 4,100 as $CaCl_2$. The researches of A. C. Lane have shown that the scanty waters in the deep levels of the copper mines near Houghton have a similar composition, except that here calcium chloride prevails. These waters, which are found in amygdaloid lava flows and associated sedimentary rocks of the Upper Algonkian (Keweenawan), are perhaps to be regarded as residual oceanic waters, which, in their long contact with the rocks, have undergone considerable changes. An analysis is given below (p. 50) This water contains no barium.

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[1] E. M. Shepard, Underground waters of Missouri, *Water-Supply Paper* 195, U. S. Geol. Survey, 1907, p. 81.

In oil bearing districts salt waters are of very frequent occurrence. They are rich in sodium chloride and often also contain the chlorides of calcium and magnesium, as well as more or less bicarbonates. They are always poor in sulphates, and this is perhaps due to their reduction by the hydrocarbons. Such salt solutions have been variously interpreted as connate waters, as solutions of saline deposits and as of magmatic origin.[1]

In the western States many similar waters occur in the Red Beds, but, as stated they are usually also rich in calcium sulphate. As an example may be cited the tepid Quelites Spring in New Mexico,[2] which ascends through Red Beds and contains about 2.6 per cent. of solids; one-half is calculated as sodium chloride and the larger part of the remainder as calcium sulphate. Bromine, boron, and barium are present. On the Pacific coast such waters are not common. Byron Hot Springs, California, may be cited as an example. The temperature spring is 76° C. The water contains about 13,000 parts of salts per million, of which over 10,000 parts are sodium chloride. A large portion of the remainder consists of calcium chloride. Small quantities of bromine, iodine, and barium are present.[3]

The Triassic strata of the French Alps and the Pyrenees are rich in similar waters, many of which are warm. The mineral combination is a characteristic mingling of chlorides and sulphates, and undoubtedly all of the constituents are derived from the sedimentary rocks mentioned.

The Spring of Mey in Haute Savoie, with a temperature of 39.8° C., may be taken as a typical example. It contains both carbon dioxide and hydrogen sulphide and yields a total of 5,000 parts per million of dissolved salts, of which 1,753 parts are calculated as sodium chloride, 1,773 as sodium sulphate, and 957 as calcium sulphate. Some bromine and traces of iodine, phosphorus, and arsenic are present.[4]

A celebrated group of these chloride springs are found in Germany on both sides of the Rhine. Among them are the

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[1] C. W. Washburne, Chlorides in oil field waters, *Trans.*, Am. Inst. Min. Eng., vol. 45, 1915, pp. 687–693. G. S. Rogers, Chemical relations of the oil field waters in the San Joaquin Valley, California, Bull. 653, U. S. Geol. Survey, 1917.

[2] F. A. Jones, New Mexico mines and minerals, 1904, p. 309.

[3] Winslow Anderson, Mineral springs and health resorts of California, 1892, p. 106.

[4] Jacquot et Willm, Les eaux minerales de la France, Paris, 1894, p. 243.

waters of Soden, Homburg, Wiesbaden, Kreutznach, Kissingen, Nauheim. Most of them issue from or ascend through salt-bearing beds of Devonian, Permian, or Triassic age, and their composition is similar. The springs of Kreutznach are especially rich in calcium chloride. Some of the springs cited are hot, others cold; some are rich in carbon dioxide. In regard to Kreutznach and Wiesbaden there is room for doubt, for the former springs stand in intimate relation to eruptive rocks, while the latter issue from a gneiss and are by some authors considered of juvenile origin. The majority of them, at any rate, have certainly derived their salts from sedimentary beds.

The chloride waters, described above, are capable of dissolving and depositing many metallic substances and have strong dehydrating power. Their relation to mineral deposits will be mentioned later.

COMPOSITION OF SALTS AND TOTAL SALINITY OF CHLORIDE WATERS
(Cited from Clarke's Geochemistry, 1916, pp. 182–186)

	A	*B*	*C*	*D*	*E*	*F*
Cl	55.83	58.79	42.00	62.31	63.55	56.58
Br	0.04	trace	1.13	0.53	....	0.04
I	0.03	....	0.02	0.01	....	trace
SO_4	3.12	0.94	0.08	0.03	0.01	0.78
CO_3	2.63	0.61	18.59	0.27	0.01	3.13
B_4O_7	....	....	....	....	....	0.01
Na	33.09	30.38	27.62	18.35	5.63	32.60
K	0.27	3.76	0.78	1.55	....	1.16
Li	....	....	0.08	0.04	....	0.04
NH_4	....	....	....	0.23	....	0.07
Ca	3.72	4.90	6.03	13.86	30.78	4.05
Ba	....	....	0.09	....	....	0.01
Sr	....	....	....	....	....	0.12
Mg	1.13	0.40	3.41	2.53	0.01	0.61
Al_2O_3	....	0.02	....	0.02	....	....
Fe_2O_3	0.06	....	0.03	....	....	....
Fe	....	....	....	0.25	....	0.04
SiO_2	0.08	0.20	0.14	0.02	0.01	0.76
	100.00	100.00	100.00	100.00	100.00	100.00
Salinity, parts per million.	10,589	23,309	12,022	309,175	212,300	8,241

NOTES RELATING TO ABOVE ANALYSES

A. Cincinnati artesian well, Cincinnati, Ohio. Analysis by E. S. Wayne, cited by A. C. Peale, *Bull.* 32, U. S. Geol. Survey, 1886, p. 133. This water contains considerable quantities of free H_2S and CO_2.

B. Utah Hot Springs, 8 miles north of Ogden, Utah. Temperature 55° C. Analysis by F. W. Clarke, *Bull.* 9, U. S. Geol. Survey, 1884, p. 30.

C. Congress Spring, Saratoga, New York. Analysis by C. F. Chandler, cited by A. C. Peale, in *Bull.* 32, U. S. Geol. Survey, 1886, pp. 38, 39. Traces of F, P, B, Sr, and Al. Contains much free CO_2.

D. Brine from well 2,667 feet deep at Conneautsville, Pennsylvania. Analysis by A. E. Robinson and C. F. Mabery, *Jour.*, Am. Chem. Soc., vol. 18, 1896, p. 915. A little H_2S is present.

E. Water from the deep levels of the Quincy mine, Hancock, Michigan. Analysis by George Steiger.

F. The Kochbrunnen, Wiesbaden, Germany. Analysis by C. R. Fresenius. This water also contains traces of I, P, and As.

CHLORIDE WATERS IN IGNEOUS ROCKS

Waters rich in chlorine are sometimes found as ascending springs in igneous rocks, but almost always close to regions of comparatively recent volcanic activity. Their composition is somewhat different from the brines resulting from the dissolving of salts from sedimentary beds. Bromine is seldom present except in mere traces, while boron appears in considerable amounts. Such tepid salt waters arise, for instance, in the volcanic region around Clifton, Arizona. The Paleozoic rocks of this region are not known to contain either salt or gypsum. Another case is the Glenwood Hot Springs in western Colorado; the springs at this place issue from limestone, but the structural relations show that the basal granite underlies this limestone at slight depths. The temperature is 49.5° C; the water contains a large amount of sodium chloride and relatively small amounts of carbonates and sulphates. Hydrogen sulphide and free carbon dioxide are present. Still another case is Steamboat Springs, Nevada, which issue from granodiorite near the eastern base of the Sierra Nevada in a region of Tertiary volcanism.

Many of these springs are rich in carbon dioxide and hydrogen sulphide; they often contain many of the rarer elements, as shown in the analyses quoted below, and they usually appear in regions rich in ore deposits. Doubt as to the derivation of the salt may exist in many cases, as, for instance, in the springs of Kreutznach, Germany, which issue from a porphyry said by

Laspeyres to contain 0.001 per cent. sodium chloride.[1] Delkeskamp,[2] on the other hand, holds that the salt is derived from sedimentary deposits. Another notable instance of chloride springs of this class is mentioned by Daubrée[3] from the provinces of Antioquia and Cauca in Columbia, where they issue in great abundance from granite, crystalline schist, and late volcanic rocks. Great difficulties arise in attempting to trace the origin of the sodium chloride in springs of this class to surrounding rocks, even admitting that granite and other crystalline rocks may contain traces of this salt. Sinters of calcium carbonate and silica are often deposited at the orifices of these springs.

COMPOSITION OF SALTS AND TOTAL SALINITY OF SODIUM CHLORIDE AND SILICA WATERS

(After Clarke's Geochemistry, 1916, pp. 186 and 196)

	A	*B*	*C*	*D*	*E*
Cl	35.00	36.61	31.64	13.52	37.52
Br	....	....	0.25	....	....
SO_4	4.58	1.84	1.30	9.01	4.96
S	0.22	....	....	0.32	....
CO_3	5.08	0.15	8.78	10.16	....
PO_4	0.03	0.08	....	....	....
AsO_4	....	....	0.24	....	....
B_4O_7	8.88	2.24	1.19	....	....
Na	30.35	21.44	26.42	19.71	24.22
K	3.79	4.45	1.93	1.88	0.36
Li	0.27	0.22	0.40	....	....
NH_4	....	0.02	trace	0.28	....
Ca	0.25	0.39	0.11	....	2.59
Mg	0.01	0.08	0.04	0.08	0.19
Fe	....	trace	trace	....	trace
As	0.10	....	....	....	....
Sb	0.02	....	....	....	....
Al_2O_3	0.01	0.76	0.12	....	0.35
SiO_2	11.41	31.72	27.58	45.04	29.81
	100.00	100.00	100.00	100.00	100.00
Salinity, parts per million	2,850	1,830	1,388	1,131	2,735

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[1] *Zeitschr.* Deutsch. geol. Gesell., vol. 19, p. 854, and vol. 20, p. 155.

[2] *Verhandl.* Polytech. Gesell. (Berlin), 1903, II, p. 161.

[3] Les eaux souterraines, etc., II, 2, p. 106.

A. Steamboat Springs, Nev. Analysis by W. H. Melville, given by G. F. Becker in *Mon.* 13, U. S. Geol. Survey, 1888, p. 349. Bicarbonate reduced to normal salts. Temperature 85° C. Contains free carbon dioxide and hydrogen sulphide. Traces of iron and quicksilver; deposits cinnabar and stibnite.

B. Coral Spring, Norris Basin, Yellowstone National Park. Analysis by F. A. Gooch and J. E. Whitfield, *Bull.* 47, U. S. Geol. Survey, 1888. Temperature 73° C. H_2S none. Free CO_2 42.5 parts per million.

C. Old Faithful Geyser. Same locality and analysts. Temperature 84°–88° C. H_2S, 0.2 part per million.

D. Great Geyser, Iceland. Analysis by F. Sandberger.

E. Water of the pink terrace, Roturoa geyser. Analysis by W. Skey.

Closely related to this group are the predominant springs in the great geyser regions of Yellowstone National Park, New Zealand, and Iceland. They are essentially sodium chloride waters with large amounts of silica, believed to exist in part as sodium silicate, a large quantity of free carbon dioxide, and a little hydrogen sulphide. Large amounts of boron, usually calculated as sodium borate, are often present, and also frequently arsenic. Bromine is rarely recorded in quantities approaching those in the brines from sedimentary formations. The waters are always hot and usually ascend through volcanic rocks, mostly rhyolite; from these the silica is supposed to be derived, but no such explanation seems sufficient to account for the predominating salt, sodium chloride, or for the boron. In the Yellowstone Park a number of the springs issuing near limestone bear evidence of their passage through this rock in increased quantities of calcium and magnesium. Others are rich in sulphate of sodium and other sulphates, but these springs give an acid reaction and the sulphates are in all probability due to the oxidation of hydrogen sulphide and the replacement of silica in sodium silicate by sulphuric acid.

SULPHATE WATERS IN SEDIMENTARY ROCKS

The waters which traverse sedimentary rocks are often rich in salts, particularly in sulphates. The gypsum waters have been mentioned and are connected with the sodium chloride waters in a manner corresponding to the association of gypsum and rock salt. By interaction of calcium sulphate and magnesium carbonate, the sulphate of magnesium may be formed, or it may be derived from the decomposition of a pyritic dolomite.

Sodium sulphate waters are almost characteristic of certain formations in the western Cretaceous, for instance; these formations consist mainly of sandstones and carbonaceous shales, the latter often pyritiferous and the whole series mainly a product of near-shore deposition. The oxidation of the pyrite furnishes solutions containing free sulphuric acid, and by reaction between this and various other substances sulphates of calcium, magnesium, and sodium will be formed. In land deposits contained in many series of sedimentary rocks sodium carbonate and sodium sulphate are formed by several well-established reactions, and percolating waters will easily abstract these salts. The interaction of calcium sulphate and sodium carbonate results in sodium sulphate and precipitation of calcium carbonate. Sodium sulphate in the presence of free carbon dioxide will dissolve calcium carbonate, forming sodium bicarbonate and a precipitate of gypsum.[1]

Reactions in soils between sodium chloride and calcium sulphate result, according to Cameron, in calcium chloride and sodium sulphate, and similar reactions take place between sodium chloride and calcium carbonate. Sodium sulphate waters are, as stated, common in the western Cretaceous, especially in the shale formations. The lowest member of this series, the Dakota sandstone, is particularly noted as a water-carrying formation. The water, which is under artesian pressure, penetrates this formation for several hundred miles underground from its outcrop and in places contains so much sodium sulphate as to be unfit for irrigation purposes. There is no evidence that this water has formed mineral deposits in the sandstone.

A well 1,400 feet deep, in Dakota sandstone at Pueblo, Colorado, contains, according to Darton,[2] 1,337 parts per million of total solids, of which about one-half is calculated as sodium sulphate and one-fourth as calcium sulphate. Very little silica and chlorine, but a little iron and carbon dioxide are present. This analysis appears to be typical. In many waters in sedimentary formations chlorides and earthy carbonates appear mixed with sulphates. Waters from artesian wells at Roswell, New Mexico,

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[1] F. W. Clarke, Geochemistry, 1916, p. 241.
E. W. Hilgard, *Am. Jour. Sci.*, 4th ser., vol. 2, 1896, p. 100.
Cameron and Bell, *Bull.* 33, Bureau of Soils, 1906.
F. K. Cameron, *Bull.* 17, Bureau of Soils.

[2] N. H. Darton, *Prof. Paper* 32, U. S. Geol. Survey, 1905, p. 355.

about 400 feet deep, derived from Permian limestones,[1] have a temperature of 64°–70° F. and contain from 600 to 1,200 parts per million of solid salts, of which 300 to 576 are calcium and magnesium sulphates and the remainder carbonates and chloride of sodium. In regions of dislocations such waters may be hot and then the ordinarily low percentage of silica may increase considerably. The Arrowhead Spring[2] of San Bernardino Valley in southern California, issuing from Tertiary sediments, has a temperature of 184° F. and contains 1,086 parts per million of solids, of which 735 are calculated as sodium sulphate, 69 as potassium sulphate, 23 as calcium sulphate, 3 as magnesium sulphate, 141 as sodium chloride, and 23 as calcium carbonate; 85 are present as silica. In well waters of the same valley the solids range from 191 to 260 parts per million and the relation is $CaCO_3 > MgCO_3 = NaSO_4 > NaCO_3 > > NaCl$. Silica amounts to 24 to 32 parts per million.

PERCENTAGE COMPOSITION AND SALINITY OF SULPHATE WATERS DERIVED FROM SEDIMENTARY FORMATIONS

	A	*B*	*C*	*D*
H_2SO_4 (free)	....	....	....	9.37
Cl	0.48	11.10	....	0.32
SO_4	66.28	59.68	76.57	68.21
CO_3	0.60	1.67	....	....
Na	30.46	13.89	1.19	0.22
K	1.08	0.49	....	0.11
Ca	0.67	2.91	5.82	0.38
Mg	0.41	10.19	3.39	1.11
Fe''	....	....	4.28	1.19
Al	....	....	7.36	11.08
SiO_2	0.02	0.07	1.39	7.11
	100.00	100.00	100.00	99.10
Salinity, parts per million	74,733	15,682	3,303	464

A. Abilena Well, Abilene, Kansas, 130 feet deep. Analysis by E. H. S. Bailey, Geol. Survey, Kansas, vol. 7, 1902, p. 166. In Permian strata. From Clarke's Geochemistry, 1916, p. 187.

[1] C. A. Fisher, Report on the Roswell artesian area, *Water-Supply Paper* 158, U. S. Geol. Survey, 1906.

[2] W. C. Mendenhall, Hydrology of San Bernardino Valley, *Water-Supply Paper* 142, U. S. Geol. Survey, 1905.

B. King's Mineral Spring near Dallas, Indiana. *Twenty-sixth Annual Report,* Indiana Dept. Geol., 1901, p. 32. Traces of Al, Fe, Ba, Sr, Li, Mn, Ni, Zn, Br, PO_4 and B_4O_7. Geological horizon Paleozoic shale.

C. Alum Well, Versailles, Missouri. Analysis by P. Schweitzer, Geol. Surv. Missouri, vol. 3, 1892, p. 131. In Pennsylvanian shale.

D. Rockbridge Alum Springs, Virginia. Analysis by M. B. Hardin. Cited in Clarke's Geochemistry, p. 98. From pyritic shale. Contains also 0.69 Mn, 0.01 Li, 0.05 Co, 0.07 Ni, 0.08 Zn, and traces of Cu, HNO_3, and PO_4.

Waters percolating through oxidizing pyritic shales sometimes contain large amounts of the sulphates of aluminum and ferrous iron; evidently this happens only when comparatively large amounts of sulphuric acid, which is capable of attacking aluminous silicates, are set free.

Such waters are not uncommon in the eastern and central States and usually contain small amounts of rarer metals; traces of nickel, zinc, and arsenic are common. The sulphate water, especially those rich in iron and aluminum, are of great importance in the genesis of deposits in the oxidizing zone, and the latter often form, at their orifices, large quantities of ocherous deposits. Many waters of this kind are known from Virginia, issuing from pyritic shales, and Peale[1] quotes some interesting and reliable analyses. A water from Alleghany Springs in Montgomery County, analyzed by Genth, contained 3,129 parts per million of solids, of which the principal constituents were calculated as 1,955 parts $CaSO_4$, 255 parts $MgSO_4$, and 61 parts $CaCO_3$. Small quantities of strontium, barium, fluorine, and silica and traces of zinc, lead, copper, and cobalt are noted. Some free carbon dioxide and a trace of hydrogen sulphide are present.

The Jordan Alum Springs in Rockbridge County, Virginia, of which several analyses by J. W. Mallet are recorded, contain from 306 to 935 parts per million of solid salts, of which the larger amount consists of aluminum sulphate, 35 to 85 parts of ferric sulphate, and from 8 to 17 parts of manganese sulphate. Small quantities of copper, zinc, cadmium, nickel, and cobalt are determined, also a trace of fluorine. One of the waters contained 102 parts of copper and 9 parts of zinc. In the Rockbridge Alum Springs, in the same State, small quantities of copper, nickel, cobalt, zinc, and a trace of lead were determined.

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[1] A. C. Peale, Lists and analyses of the mineral springs of the United States, *Bull.* 32, U. S. Geol. Survey, 1886, pp. 58–65.

Free sulphuric acid is present in the Bedford Alum Spring to the amount of 70 parts per million, according to M. B. Hardin. The total solids are 1,207 parts, practically all sulphates, and about one-third consists of ferric sulphate. Small quantities, about 0.8 part per million of each, of nickel, cobalt, copper, and zinc were determined. Springs of similar composition are found in Pennsylvania and other eastern States. All these acid springs are poor in silica and contain very little chlorine.

ACID SULPHATE WATERS IN IGNEOUS ROCKS

Sulphate springs in connection with igneous rocks and volcanism appear mainly as products of the oxidation of ascending waters of alkaline reaction, containing free hydrogen sulphide, but there is evidence that in regions of volcanic activity such oxidation takes place on a large scale and that these acid waters are of high importance in effecting rock alteration, particularly by attacking aluminum silicate and developing alunite. By similar reactions free hydrochloric acid may be generated, for instance by the decomposition of chlorides by free sulphuric acid.

As a consequence it is common to find such waters near the orifices of hot springs, as well as at volcanoes. The development of free acid of course displaces the equilibrium and the oxidized water may differ greatly from its parent liquid; thus it happens that a single ascending hot spring may yield a whole series of derivatives of varying temperature and composition by mingling with other waters and by oxidation.

A number of analyses of such waters are quoted in Clarke's Geochemistry. Some of them, especially from pools or lakes near volcanoes, are remarkably rich in hydrochloric acid. The peculiar water from the Yellowstone National Park known as the Devil's Inkpot contains, besides free acids, a large amount of sulphate of ammonia. The water from Roturoa, New Zealand, the analysis of which is quoted below, is a more characteristic product of the oxidation of normal thermal waters.

The geysers of Sonoma County, California, of which there is a good series of analyses by Dr. Winslow Anderson,[1] form a most remarkable illustration of the oxidation of hot waters. There are at this place a great number of springs of varying temperature

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[1] Winslow Anderson, Mineral springs, etc., of California, 1892, pp. 136–154

G. A. Waring, Springs of California, *Water-Supply Paper* 338, U. S. Geol. Survey, 1915, p. 109.

and composition, all of them heavily charged with hydrogen sulphide. The primary water at a temperature of 110° F. appears to contain chiefly carbonate of magnesium with some of calcium. The total solids amount to about 568 parts per million, most of which consist of the above-mentioned carbonates; there are 92 parts of silica per million. This water is probably of mixed origin; the carbonates are clearly derived from the serpentinoid rocks of the vicinity, but the hydrogen sulphide is most likely of magmatic origin. Free sulphuric acid is generated by oxidation and gives rise to a long series of peculiar sulphate waters, most of them rich in dissolved solids and of high temperature. An analysis of one of these shows 3,262 parts per million of total salts and acids, among which the sulphates of magnesium, sodium, and aluminum prevail. There are 544 parts per million of free sulphuric acid and 20 parts of free hydrochloric acid.

Finally, acid water may result directly from the oxidation of deposits of pyrite or of sulphur. A water of the latter type is described by W. T. Lee from Beaver County, Utah.

PERCENTAGE COMPOSITION OF SALTS AND TOTAL SALINITY OF ACID WATERS

	A	*B*	*C*
HCl free	0.18	5.60	
H_2SO_4 free	1.29	59.11	46.39
H_3BO_3	2.73		
Cl			0.08
SO_4	67.66	20.21	32.63
Na	0.73	8.35	1.48
K	0.24	0.32	
Li	0.01		
NH_4	22.85		
Ca	1.18	0.47	1.63
Mg	0.36	0.22	2.50
Fe'' [1]	trace		5.76
Fe'''		0.33	8.25
Al	0.10	trace	
SiO_2	2.67	5.39	1.28
	100.00	100.00	100.00
Salts and acids, parts per million	3,365	1,862	9,716

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[1] Includes some alumina.

A. Devil's Inkpot, Yellowstone National Park. Analysis by F. A. Gooch and J. E. Whitfield, *Bull.* 47, U. S. Geol. Survey, 1888, p. 80. Contains also 65 parts of free CO_2 and 5 parts of H_2S per million. Cited in Clarke's Geochemistry, 1916, p. 199.

B. Cameron's Bath, Roturoa geyser district, New Zealand. Analysis by W. Skey, *Trans.*, New Zealand Institute, vol. 10, 1877, p. 423. Contains 6 parts per million of H_2S. Cited in Clarke's Geochemistry, p. 200.

C. Water at Sulphur mine of Cove Creek, Beaver Valley, Utah. Analysis by W. M. Barr, *Water-Supply Paper* 217, U. S. Geol. Survey, 1908, p. 20. Contains also much free H_2S.

MINE WATERS OF SULPHATE TYPE

Mine waters consist as a rule of the normal surface waters of the rock containing the ore deposit, modified by the salts resulting from the decomposition of the minerals of the deposit. In deposits free from sulphides, such as the copper and iron mines of Lake Superior, there is little difference between the mine waters of the upper levels and the normal surface waters of low salinity; both are comparatively high in silica and calcium carbonate. Where much pyrite or marcasite is present, as in coal mines and in most metal mines, the surface waters will contain sulphates of ferrous iron and aluminum and frequently also of the rarer metals. When these waters mingle with normal surface waters rich in calcium carbonate the iron and alumina may be precipitated as hydroxides and calcium sulphate remains in solution. Calcium sulphate waters often spread over a considerable area surrounding pyritic deposits.

The mine waters will be discussed in the chapter on oxidation and secondary sulphides.

SODIUM CARBONATE WATERS IN SEDIMENTARY ROCKS

Waters containing sodium carbonate in large amounts are not common in sedimentary rocks, but here and there wells or springs of this character are encountered; they are usually cold and often contain some free carbon dioxide and hydrogen sulphide. The alkaline carbonate is probably, as suggested above, derived from a reaction between sodium sulphate and calcium carbonate or between sodium chloride and calcium carbonate. Waters of this kind occur at a few places in the eastern and central States. G. L. Cumming has described such waters from some artesian wells in Silurian limestone on the Island of Montreal.[1] They contain from 500 to 700 parts per million of solids, chiefly sodium

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[1] *Mem.* 72, Geol. Survey Canada, 1915.

carbonate with the remainder calculated as calcium chloride and sodium sulphate. Cumming shows that the waters are of complicated origin but believes that the sodium carbonate solutions are derived from dikes and intrusions in the limestone. A good instance is furnished by some Missouri waters in Carboniferous limestone, one of which is quoted under *E* in the following table. Similar waters are those of the wells at La Junta, Denver, and Greeley, Colorado. The artesian wells at Denver, about 1,200 feet deep, are in the Arapahoe Eocene, while the Greeley well, of the same depth, is sunk in Laramie sandstone. The maximum of total solids is about 1,530, divided between sodium carbonate and sodium chloride. Some free CO_2 is present. Many artesian waters in New South Wales are rich in sodium carbonate. Sedimentary beds containing volcanic tuffs often yield sodium carbonate waters.

SODIUM CARBONATE WATERS IN IGNEOUS ROCKS

Ascending sodium carbonate waters are most characteristic of regions of subsiding or expiring volcanism. During surface eruptions alkaline chlorides and carbonates always appear as sublimates and waters traversing tuffs, breccias and lava flows may dissolve these salts together with other volcanic exhalations such as borates. The characteristic sodium carbonate waters are, however, of deep-seated origin and usually break in through the older igneous or metamorphic rock underlying the lavas in regions where the active volcanism has ceased, and the prevailing opinion is that these waters with their charge are in whole or in part of magmatic origin. They rarely contain much calcium and they are poor in silica, but are usually heavily charged with carbon dioxide and sometimes hydrogen sulphide. They almost always contain many rarer substances such as boron, fluorine, iodine, arsenic and various metals.

Such waters would attack silicates with great energy and it is suggested that their strong percentage of sodium may have been leached from walls of the fissures, during the conversion of sodium silicates to potassium silicates as in the process of sericitization, so common in mineral veins. It is certain that these waters are of the utmost importance in the genesis of orebearing veins.

An excellent instance of a province of such waters is furnished by the volcanic district of central France.[1] An analy-

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[1] Jacquot and Willm, Les eaux minerales de la France, Paris, 1894.

sis of the celebrated Vichy Springs is given in the table on this page. Sodium preponderates as bicarbonate, but smaller quantities of sodium chloride and sulphate are also present. The whole region of the Central Plateau is rich in carbon dioxide, occurring both in springs and as exhalations (for instance, at the Pontgibaud lead-silver mines). The magmatic source of the gas is rarely questioned, whatever opinion may be held about the origin of the water (Fig. 4).

In the volcanic regions of Taunus and Vogelsgebirge on the Rhine in Germany are the springs of Ems and Fachingen. The springs of Ems issue with a temperature of 46° C. and contain about 2,870 parts per million of solids, of which about one-half may be calculated at sodium carbonate and a large part of the remainder as sodium chloride.

PERCENTAGE COMPOSITION OF SALTS AND TOTAL SALINITY OF SODIUM CARBONATE WATERS

(Cited from Clarke's Geochemistry, pp. 191, 193 and 197)

	A	*B*	*C*	*D*	*E*	*F*
Cl	6.17	8.85	11.52	13.57	6.63	4.01
F	....	0.19	0.03	....	....	....
SO_4	3.75	5.77	31.19	0.32	6.21	4.26
S	....	....	....	....	0.06	....
CO_3	45.57	41.91	19.15	22.38	44.76	47.45
PO_4	1.52	0.01	0.01	....	....	....
AsO_4	0.04	....	....	....	....	....
B_4O_7	....	0.16	....	27.98	....	....
Na	35.27	38.08	32.49	33.97	41.07	40.09
K	2.88	1.20	1.35	0.48	....	0.38
NH_4	....	....	....	0.05	....	....
Li	....	0.12	....	....	....	....
Ca	2.29	0.87	2.23	0.41	0.30	0.27
Sr	0.04	0.05	0.01	....	....	....
Mg	1.11	0.41	0.65	0.11	0.12	0.15
Mn	....	....	0.01	....	....	....
Fe	....	....	0.02	....	....	0.14
Fe_2O_3	0.04	0.06	....	....	....	....
Al_2O_3	....	0.02	....	....	....	0.20
SiO_2	1.32	2.30	1.34	0.73	0.85	3.05
	100.00	100.00	100.00	100.00	100.00	100.00
Salinity, parts per million	5,249	2,614	5,431	5,096	2,069	1,668

NOTES RELATING TO ABOVE ANALYSES

A. The Grand-Grille spring, Vichy, France. Analysis by J. Bouquet. Small quantity of fluorine present. Temperature 44° C. Issues from Tertiary beds.

B. Ojo Caliente spring, near Taos, New Mexico. Analysis by W. F. Hillebrand. Trace of barium and arsenic. In lake beds and gneiss.

C. The Sprudel, Carlsbad, Bohemia. Analysis by F. Ragzsky. Contains 0.76 gram free and half-combined CO_2 per kilogram. Traces of Br, I, Li, B, Rb, and Cs. Temperature 72° C. In granite.

D. Hot water from the Hermann shaft, Sulphur Bank, California. Analysis by W. H. Melville. A little H_2S and a considerable amount of CO_2 present. Temperature 80° C. In basalt and sandstone.

E. McClelland well, Cass County, Missouri, 45 feet deep, in Carboniferous limestone. Analysis by P. Schweitzer. Contains H_2S.

F. Artesian water, La Junta, Colorado. Well 386 feet deep. Analysis by W. F. Hillebrand. In Cretaceous beds.

At the foot of the Erzgebirge, in the Tertiary volcanic region of northern Bohemia, issue a series of hot springs, extending from Teplitz to Carlsbad and Eger. Most of these belong to the class of sodium carbonate waters with free carbon dioxide. They contain an abundance of salts, and in the Teplitz and Bilin springs sodium carbonate predominates. In the Carlsbad (*C* in table of analyses) and Marienbad spring the sulphuric acid radicle is prominent and must largely exist in sodium sulphate. The Carlsbad springs contain fluorine and barium with traces of many rarer metals which are mentioned on page 97.

In the Cordilleran Ranges in North America and South America sodium carbonate waters are abundant and always closely connected with areas of Tertiary volcanic activity.

In New Mexico the Ojo Caliente (*B* in table of analyses), Faywood, and Las Vegas springs may be mentioned; in Colorado the Idaho Springs, Middle Park Springs, Poncha Springs, and the water in the Geyser mine at Silver Cliff; in Idaho the Boise Hot Springs. In California sodium carbonate waters are especially abundant and characteristic; they follow the Coast Range from San Diego to Mendocino County and appear to stand in some causal connection with the late Teritary or Quaternary eruptions of basalt. Some of the waters are clearly admixed with magnesium from the serpentinoid rocks which they have traversed, but in general the type is perfectly distinct. The following data are taken from the U. S. Geological Survey, *Bulletin* 32, by A. C. Peale.

Source of water	Salinity, parts per million	Composition and quantity of principal salts
San Juan Capistrano (T. 50° C.)	290	$HNaCO_3 > NaCl > SiO_2$ 111 105 70
Skaggs Springs (T. 54° C.)	2,556	$HNaCO_3 > BO_2 > SiO_2$ 2,083 176 151
Paso Robles Springs (T. 42° C.)	1,581	$HNaCO_3 > NaCl > Na_2SO_4$ 850 469 136
New Almaden Vichy (T. 17° C.)	7,361	$HNaCO_3 > CaSO_4 > CaCO_3$ 3,400 680 544
Napa Soda (T. 17° C.)	1,156	$HNaCO_3 > MgCO_3 > NaCl$ 561 187 85
Pacific Congress (T. 10° C.)	5,678	$HNaCO_3 > NaCl > CaCO_3$ 2,091 1,923 289
Ukiah Vichy (T. 34° C.)	4,624	$HNaCO_3 > NaCl > MgCO_3$ 3,369 459 374

An interesting type of these waters is represented by the hot spring of Sulphur Bank (*D* in table of analyses), which contains boron and is depositing cinnabar. On the whole these waters are rich in unusual constituents and have great solvent powers.

SODIUM SULPHIDE WATERS

It is believed that in some of the springs already referred to—for example, Steamboat Springs, Nevada—sodium sulphide or other sulphur salts of sodium are present. In the Pyrenees of France and Spain is found a group of springs in which sodium sulphide is constantly present. These springs have a high temperature and a low salinity, containing from 250 to 350 parts per million of salts; they usually issue in crystalline schists or on the contact of the schists with Paleozoic strata. A characteristic spring mentioned among others by Jacquot and Willm[1] contains total CO_2 52, S (in sulphides) 31, Na 97, CO_3 26, Cl 55, and SiO_2 93 parts per million. Some organic matter is present and strong traces of boron, arsenic, copper, etc., are mentioned. There appears to be considerable difficulty in the explanation of this combination on the hypothesis of leaching from the surrounding country rock. Where contaminated by surface water

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[1] Lea eaux minerales de la France. Paris, 1894.

or where locally issuing through Triassic strata they beocme calcic. By oxidation they acquire hyposulphites.

SUMMARY

In sedimentary formations, beyond the influences of igneous activity, the waters are of many differing types. Some contain mainly calcium carbonate; others are of the chloride type, with sodium or calcium as the prevalent base; still others, a very abundant class, are rich in calcium or sodium sulphates; a rarer type is that of the sodium carbonate waters. Naturally many waters show a mingling of these types. Most of these waters are cold; many are tepid; few of them are hot. Whether warm or cold, both hydrogen sulphide and carbon dioxide may be present.

In older igneous rocks where the effects of volcanism have subsided the types vary less widely. The ordinary surface waters are always—unless some disturbing influence interferes—of the calcium carbonate type, often with sodium chloride, ferrous and magnesium carbonate, and considerable silica, but low salinity. These waters sometimes, but not often, appear as tepid ascending springs. If the rocks contain iron disulphide the waters may locally contain free sulphuric acid and the sulphates of calcium, aluminum and iron.

The remaining classes of water in igneous rocks are ascending and confined to regions of recent or Tertiary volcanic activity. They are tepid to hot, though cold waters are also known. They easily fall into two classes: (1) the sodium chloride waters, of which the siliceous "geyser waters" form a sub-class; (2) the sodium carbonate waters, which are generally rich in free carbon dioxide. Transitions between the two classes are plentiful, and the latter class may in rarer cases also contain notable amounts of sodium sulphate; of this class the Carlsbad Springs form a prominent example.

INTERPRETATION OF WATER ANALYSES

Analyses of waters are usually stated in parts per million of radicles and metals. From this form a calculation will be necessary to ascertain whether the water is alkaline, neutral, or acid.

Stabler[1] has suggested that for this purpose the quantities deter-

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[1] Herman Stabler, The mineral analysis of water for industrial purposes and its interpretation by the engineer: *Eng. News*, vol. 60, 1908, p. 356. Also, chapter on the industrial application of water analyses in *Water-Supply Paper* 274, U. S. Geol. Survey, 1911, pp. 165–181.

mined may be multiplied by the reciprocals of the equivalents. The products are called the reacting values. If the water is neutral the reacting values of acids and basic radicles should balance. Palmer,[1] in his method of geological interpretation of water analyses, finds it convenient to express the reacting values in percentages, thus eliminating the factor of concentration. Palmer's classification emphasizes the fact that a solution in which strong acids are exactly balanced with strong bases is relatively inert, whereas one in which either group exceeds the other is relatively active. It is of special use in showing the relationship and the nature of chemical action of different waters.

Alkalinity and *Salinity* are the fundamental properties. Salinity is measured by the strong acid radicles (SO_4, Cl). If the basic radicles are partly or wholly alkaline metals their proportion of the salinity is said to be *primary*. The remaining salinity due to radicles Ca, Mg, Fe is called *secondary*. If the acid radicles are in excess, *tertiary* salinity or acidity results. The measure of *primary alkalinity* is the excess of alkaline metal radicles over the strong acids; the weak-acid radicles CO_3 and HCO_3 which balance any excess of the alkaline earth metals over the stronger acids produce *secondary alkalinity.*[2]

In spite of an objectionable terminology Palmer's method furnishes a convenient basis for comparative study but as a classification of natural waters it is unwieldy and uncertain. It is not always a safe guide to the geological history of the water.

The constants used in converting grains per gallon to parts per million and vice versa are as follows:

1 grain per U. S. gallon = 17.138 parts per million
1 grain per Imperial gallon = 14.285 parts per million
1 part per million = 0.0588 grain per U. S. gallon
1 part per million = 0.07 grain per Imperial gallon

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[1] Chase Palmer, Geochemical interpretation of water analyses, *Bull.* 479, U. S. Geol. Survey, 1911.

[2] Cfr. F. W. Clarke, Geochemistry, 1916, p. 63.

G. S. Rogers, The interpretation of water analyses by the geologist. *Econ. Geol.*, vol. 12, 1917, pp. 56–88.

CHAPTER V

THE CHEMICAL WORK OF UNDERGROUND WATER

METAMORPHISM AND MINERAL DEPOSITS

Stability of Minerals and Rocks.—The underground water plays a very important part in the changes which take place in rocks, and the majority of mineral deposits are formed by the aid of it. Near the surface it may completely saturate the rocks or move in large volumes on fractures. At greater depths where there is no active circulation it may be sparingly present as rock moisture. The great mass of underground water is of atmospheric origin but as all magmas contain water which is given off upon solidification some waters in the rocks may be of magmatic origin. Solution and precipitation go on continuously; one or the other may predominate at any given place. The reactions which take place in the underground solutions extend over a wide range as to temperature, pressure, substances, concentration and time, and they differ markedly under the varying conditions. The study of these reactions was first seriously undertaken by G. Bischof and Justus Roth[1] and these pioneers have been followed by many eminent geologists who have devoted themselves to the study of chemical geology.

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[1] G. Bischof, Lehrbuch der chemischen und physikalischen Geologie, 1863-1866.

Justus Roth, Allgemeine und chemische Geologie, vol. 1, Berlin, 1879.

C. R. Van Hise, A treatise on metamorphism, *Mon.* 47, U. S. Geol. Survey, 1904.

C. R. Van Hise, Metamorphism of rocks and rock flowage, *Bull.*, Geol. Soc. America, vol. 9, 1898, pp. 269–328.

C. R. Van Hise, Some principles controlling the deposition of ores, *Trans.*, Am. Inst. Min. Eng., vol. 30, 1900, pp. 27–177.

U. Grubenmann, Die krystallinen Schiefer, Berlin, 1910.

F. Becke, Ueber Mineralbestand und Struktur der krystallinen Schiefer, Ninth Session Internat. Geol. Congress, Vienna, 1903; also Sitz. Ber., k. k. Akad., Vienna, 1903.

John Johnston and Paul Niggli, The general principles underlying metamorphic processes, *Jour. Geology*, vol. 21, 1913, pp. 481–516; 588–624.

C. K. Leith and W. J. Mead, Metamorphic Geology, New York, 1915.

One of the most fruitful conceptions developed in recent years is that of the limits of stability of minerals and rocks. Conforming to increasing heat and pressure, zones exist in the earth's crust, gradually merging into one another but each characterized by certain groups of minerals that are stable only under the conditions prevailing in that particular zone. No mineral is absolutely stable. If subjected to certain conditions of temperature or in contact with certain solutions it will melt, decompose, dissociate or dissolve. At the surface under the influence of atmospheric waters with oxygen and carbon dioxide practically no minerals are stable except a few oxides, hydroxides and native elements.

In consequence of the reversible nature of chemical processes under changing conditions each mineral thus has its stability field or "critical level" which it can not leave without undergoing decomposition. The mineral aggregates, that is, the rocks, also follow this law and as the rock minerals have usually been formed in closely analogous ways most of the component minerals will become unstable more or less simultaneously.

Certain minerals, few in number, are less sensitive than others to such changes and recur under the most different conditions. They are designated "*persistent minerals*" and are in general of simple composition and do not contain the hydroxyl molecule; among them are quartz, magnetite, pyrite, chalcopyrite, fluorite, calcite and native gold. Orthoclase, all plagioclases, biotite, augite, olivine, the spinels, cordierite, and garnets develop and are fully stable only at high temperatures. Minerals rich in water, like chlorite, serpentine, and talc, are characteristic of lower temperatures. Other minerals, like muscovite, zoisite, epidote, hornblende, and albite, develop by preference under strong pressure.

The varied composition of the crust, the unequal distribution of the underground water, the changing pressure, and the great differences in temperatures even at the same horizon make it difficult to establish strict rules and well-defined zones. One merges into another. Besides, stability is a relative term. Some rocks, like granite, are really stable only shortly after their complete consolidation. Under the influence of percolating deep waters the minerals of the granite are unstable, as they are in the zone of weathering. But the changes take place so slowly that at many places they can scarcely be perceived. Other rocks, like calcareous shales, are stable at moderate depths, but easily subject to recrystallization under pressure and rising temperature. The results of the reactions differ widely according to the composition of the waters. The minerals that develop in a rock charged with a slight amount of moisture are not the same as those that appear when the rock is penetrated by rapidly moving solutions, charged with salts and gases of foreign origin.

Metamorphism.[1]—The term *metamorphism* meaning strictly "a change in form," was proposed by Lyell in 1833 to express the changes of sedimentary beds to slates, quartzite, crystalline limestone, etc. Later it was extended to the development of schists and slates from igneous rocks by pressure and recrystallization. Still later, for instance, by C. R. Van Hise it has been employed in a wide sense so as to cover any change in the composition and structure of any rock, through whatever agency and with or without gain or loss of substance. This would include weathering and the development of any kind of epigenetic deposit, such as mineral veins, in a rock. Geologists have not generally accepted this wide definition. Metamorphism is here reserved for the processes which result in a partial or complete crystallization or recrystallization of solid masses of rocks,[2] as in gneiss from granite or mica schist from clay shale. Though the mechanical effects of pressure may be conspicuous, metamorphism is always characterized by chemical changes in the component minerals. The composition of the rock as a whole may remain fairly constant.

For practical purposes we may distinguish between *static*, *dynamic*, *igneous* and *hydrothermal* metamorphism. Static metamorphism proceeds without stress, at slight depths and under influence of a slight amount of water. At great depths and high temperatures a static recrystallization under great load may be recognized.[3] Dynamic metamorphism is effected under stress at higher or lower temperatures. These two are regional and proceed without marked changes in composition.

Igneous metamorphism includes the effects of magmas on adjacent rocks and is a high temperature process. It is about

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[1] For a thorough discussion of the various uses of this term, see: R. A. Daly, Metamorphism and its phases, *Bull.* Geol. Soc. Am., vol. 28, 1917, pp. 375–418.

[2] A. Harker, *Geol. Mag.*, vol. 6, 1889, p. 15.

[3] R. A. Daly, *op. cit.*, p. 400.

equivalent to contact metamorphism but includes also the effect of igneous injection and pegmatitization.

Hydrothermal metamorphism includes the changes effected in rocks by circulating hot ascending waters. Igneous metamorphism may be local or regional and in part involves changes of composition. Hydrothermal metamorphism is local and almost always involves changes of composition.

Metasomatism or Replacement.—The geological importance of metasomatism or replacement has already been pointed out on p. 26. The word metasomatism, meaning a change of body, first used by C. Naumann to designate some kinds of pseudomorphism, is now applied to the process of practically simultaneous capillary solution and deposition by which a new mineral of partly or wholly differing chemical composition may grow in the body of an old mineral or mineral aggregate. The secondary minerals of any metamorphic rock result from metasomatic action. Rocks are termed metasomatic if their composition has been materially changed by replacement of the original minerals. Pseudomorphs and petrifications often furnish direct and incontrovertible evidence of processes of replacement.

Metasomatism is met everywhere and at all depths in sedimentary and igneous rocks and shows that the rock minerals have been subjected to conditions under which they were unstable. The development of chlorite in augite, sericite or kaolin or calcite in feldspars, or galena in limestone is due to metasomatism. The typical metasomatic processes, traced with the highest magnifying power, show no space between the parent mineral and the metasome, as the newly developed mineral may be designated. The fibers and blades of sericite project into quartz without the slightest break in the contact. Rhombohedrons of siderite develop in quartzite, their crystal faces cutting across the grains without any interstices. Perfect prisms of tourmaline develop in feldspar grains, and sharp cubes of pyrite in primary feldspar or quartz.

Metasomatic rocks, that is rocks which have suffered a change in composition, are very common in mineral deposits and are often produced by strong and rapidly moving solutions (usually aqueous, sometimes gaseous) which penetrate the material through veinlets and pores. There are many cases of complete or almost complete metasomatism, for instance of limestone by sulphides and quartz in which the chemical composition has been absolutely changed. In contrast to this the ordinary metamorphic processes in rocks are carried on by the scant rock moisture and while there is metasomatism in detail, the composition as a whole is but little changed. For description of metasomatic processes and for criteria of metasomatism see Chapter XI.

Dissemination refers to grains or crystals distributed in a rock and is without genetic significance.

Impregnation is a genetic term and means that the mineral introduced is later than the rock; it may have developed by metasomatic processes or by filling of pore spaces or other cavities.

Cementation is used to indicate the filling of interstices in porous or shattered rocks.

The Law of Equal Volume.—It is necessary to distinguish between (1) metasomatic changes proceeding in free crystals or grains, or in loose aggregates under light load, where the force of crystallization can easily overcome the restraining pressure, and (2) metasomatic changes proceeding in rigid rocks where the new mineral is forced to make room for itself by solution of the host mineral.

In the first case the volume changes proceed according to the chemical formula. In the second case, *the replacing mineral occupies exactly the space formerly filled by the primary mineral;* the force of crystallization is of little or no direct influence, but as the pressure differs in intensity according to the crystallographic directions and as solution proceeds most actively at points of greatest pressure the development of crystal faces is thereby explained.[1]

The chemical formulas by which some kinds of replacement are usually expressed do not represent the actual change for these formulas are based on equal weights and will indicate definite changes in volume. The conversion of orthoclase to sericite is usually considered to take place according to the following reaction,

$$\underset{\text{Orthoclase}}{3KAlSi_3O_8} + H_2O + CO_2 = \underset{\text{Sericite}}{KH_2Al_3Si_3O_{12}} + K_2CO_3 + 6SiO_2$$

which involves a decrease in volume of 15.5 per cent. even if the SiO_2 is assumed to have recrystallized as quartz. If, how-

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[1] W. Lindgren, The nature of replacement, *Econ. Geol.*, vol. 7, 1912, pp. 521–535. Volume changes in metamorphism, *Jour. Geol.*, vol. 26, 1918, pp. 542–555.

ever, one volume of orthoclase has been replaced by an equal volume of sericite this equation is not correct, and by a calculation of the quantities of silica, alumina, etc., contained in one cubic centimeter of orthoclase and sericite, respectively, it will be found that a considerable addition of alumina is necessary. The actual formula is probably very complicated and could be established only if all the reactions taking place in the solution during the conversion of one mineral to the other were known. Many kinds of metasomatism, for instance, galena or barite replacing calcite (Figs. 62 and 64) can not be expressed by chemical formulas. One crystal, for instance, of pyrite may simultaneously replace parts of adjacent grains of different minerals, or may replace an aggregate of minerals in a fine grained rock. These well known facts will at once show that replacement is not the expression of one definite chemical reaction.

The law of equal volumes has been repeatedly verified by many independent observers and there is little doubt that it holds for most metasomatic processes, both on a large and a small scale, both in general metamorphism and in mineral deposits. The most fundamental changes in rocks take place with practical constancy of volume. A great deal has been written on changes of volume and energy liberated or absorbed, that is absolutely valueless as a measure of the processes that have been going on in rocks. The time will soon come when these relations are more clearly recognized.

Metasomatism in solid rocks proceeds independently of molecular weight, molecular volume and specific gravity. It does not take place "molecule for molecule," and it is not expressible in simple chemical equations. At the same time it is molecular or at least sub-microscopic in the sense that complex processes of solution and precipitation constantly take place in the solution films.

Metasomatism by equal volume takes place in most perfect form when a rock is permeated by stagnant or slowly moving solutions. When, as sometimes happens, the solutions move rapidly, the nice equilibrium is disturbed. Local excess of solution over deposition is then expressed in drusy or cellular structure, and the law of equal volumes may fail to hold.

It is not meant that the processes of solution, precipitation and chemical reactions which make up metasomatism do not obey the law of mass action and laws of LeChatelier and Van't Hoff. They undoubtedly do. With increasing pressure a denser mineral will tend to form but the saturated solutions that fill the rock will at once proceed to fill any available space created by the deposition of the denser mineral with the next combination ready to be precipitated.

General Definition of the Metamorphic Zones.—That part of the earth's crust which is within our observation is called the lithosphere. It may be observed directly by borings or mining operations or indirectly when deformation and denudation bring up to the surface rocks which we know were once deeply buried. The lower limit of the lithosphere can, of course, not be accurately fixed; F. W. Clarke has suggested that it may be defined as extending 10 miles, or 16 kilometers, below the surface. Below the lithosphere lies the centrosphere, concerning which we have little definite information.

The conceptions of Albert Heim developed by C. R. Van Hise led to a division of the lithosphere into an upper zone of fracture and a lower zone of rock flowage, in which only subcapillary openings exist (p. 30) and deformation is effected by granulation and recrystallization. Between them intervenes a middle zone of combined fracture and flowage. The limits of these zones are very indefinite owing to the greatly differing plasticity of rocks, *e.g.*, a granite and a calcareous shale. The experimental proof given by F. D. Adams[1] that in supported rocks in depth openings in granite can persist to depths of at least 11 miles, or about 58,000 feet, at a uniform pressure of 70,000 pounds per square inch and at temperatures supposedly corresponding, that is, 550° C. shows that the zones overlap widely and have only value as relative conceptions.

Van Hise divided the zone of fracture in an upper zone of weathering and a lower zone of cementation. The zone of flowage corresponds to the deep metamorphic zones in which minerals form by replacement only and in which the temperature is high and the pressure largely stress.[2]

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[1] F. D. Adams, *Jour. Geology*, vol. 20, 1912, pp. 97–118.

[2] Van Hise called the upper two zones the realm of katamorphism and the lower that of anamorphism. In the zone of katamorphism (kata, down) complex silicates break down and simpler, less dense minerals form. In the zone of anamorphism (ana, up) silicates are supposed to be built up with forming of denser minerals and compact texture. Since Leith and

Later investigations have shown that any rock may be deformed under stress.[1] The thrust required to develop deformation in marble at a pressure corresponding to 4.2 miles would be 66,400 pounds per square inch; in case of granite, 138,500 pounds per square inch. At greater depths the required stress increases markedly. The pressure necessary for plastic deformation is very much greater than the crushing strength of the rock at the surface.

Zone of Weathering.—The best defined zone is that of weathering, the depth of which is determined by the level of the groundwater, or by the depth to which free oxygen can penetrate in large quantities. In the zone of weathering the water percolates downward more freely than in the underlying zone, there is a tendency to the destruction of the rocks as units, and active transportation and concentration are characteristics.

Chemical work progresses by means of water solutions and gases, also extensively through the medium of organic life; mechanical disintegration is also important. The chemical reactions are oxidation, carbonatization, desilication, and hydration, the two first named mainly through decomposition of silicates by water containing carbon dioxide. As a consequence of these reactions the volume should increase, but so much is carried away by solution that a great reduction of volume ensues.

Disintegration works hand in hand with decomposition and in advance of it; calcium, magnesium, sodium, and potassium are leached; the final products are a small number of minerals, largely hydrated compounds with low specific gravity and, for the most part, comparatively simple molecules. Almost all rock-forming minerals are unstable, as are the sulphides. These processes give rise to many mineral deposits of oxidized ores. which will be described in a later chapter.

The great extent of weathering and the intensity of the changes are justly emphasized, especially in regions of soluble rocks like limestone. It is well to bear in mind that this is not because of rapid attack by waters, but because of long-continued action by extremely dilute solutions. This is shown by the relative purity

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Mead have changed these conceptions (Metamorphic Geology, 1915) and now confine katamorphism to processes of weathering a confusion has been introduced that is best cured by the dropping of both terms.

[1] F. D. Adams and J. A. Bancroft, *Jour. Geology*, vol. 25, 1917, pp. 597–637.

of the surface waters, which contain calcium and magnesium carbonates with lesser amounts of alkaline salts. The soluble products mainly escape into the rivers through the zone of discharge, which lies below the zone of weathering, and finally into the oceans.

The Intermediate Zone.—The rocks immediately below the zone of weathering are often saturated with water which diminishes in quantity with increasing depth. The small pressure permits fracturing and brecciation, and the openings created by these processes, as well as those resulting from porosity, are filled with minerals deposited by circulating solutions. To a small extent these minerals result from material abstracted from the zone of weathering, but that zone is shallow in comparison with the zone of cementation, and the salts available from the weathering are, to a large extent, carried away by the surface drainage. The larger part of the minerals deposited have been derived from the rocks themselves; to a considerable extent they are derived from deep-seated sources, as, for instance, in the cementation by quartz veins and veinlets near igneous intrusions. Hydration and carbonatization are the principal processes. Minerals like chlorite, serpentine, talc, sericite, epidote, and calcite develop, largely by metasomatic processes. Replacement and filling work together.

Where stress is present it is mainly in one direction and shearing and schistosity may develop; in metamorphic schists some of the minerals formed are muscovite, chlorite, talc, hornblende, zoisite, epidote, and albite; also quartz, pyrite, and calcite, probably magnetite and specularite. The clay slates with muscovite and albite, the chloritic schists, and the talc schists belong to this zone. According to the views of Van Hise such schistose rocks can only develop in the deeper zones.

The Deeper Zones.—In the deeper belts (included by Van Hise under the name of the anamorphic zone) the pressure and temperature are high; the latter in general above 200° C. Very little water is present. Minerals are formed mainly by replacement. In the upper part of the zone temperature and pressure work in the direction of diminished molecular volume. The pressure is largely stress—that is, acting in one direction—but hydrostatic pressure (transmitted in all directions) is becoming of importance. The important reactions are dehydration, the development of silicates, and deoxidation. Heavy silicates, like wollastonite, garnet, tremolite, and diopside, form in siliceous limestones or in pure limestones where the silica is supplied by plutonic intrusions. The minerals produced are numerous, stable, heavy, and complex. The rocks formed are compact and strong. However, the temperature is not sufficiently high to break up the molecules in which hydroxyl is firmly contained.

The recrystallization takes place according to the law of Riecke,[1] so that the solution prevails at places of maximum pressure, and deposition at those of minimum pressure. The recrystallized products may assume lamellar structure extending perpendicularly to the pressure; this results in a "schistosity by crystallization." Among the minerals of this zone are muscovite, microcline, albite, microperthite, oligoclase, biotite, zoisite, epidote, hornblende, staurolite, garnet, cyanite, titanite, magnetite, and ilmenite. Most of the micaceous and hornblendic gneisses containing garnet, staurolite, etc., belong to this zone; also the mica schists, amphibolites, and glaucophane rocks.

Where there is no stress in this zone, many igneous rocks, like granite, basalt, and rhyolite, are stable.

In the lower part of the deep-seated zone the temperature is high and the tendency is toward an increase of volume. The hydrostatic pressure is enormous and stress almost non-existent, but high temperature is the dominant feature. There are no minerals containing the hydroxyl molecule except biotite, and the characteristics are, therefore, the prevalence of anhydrous minerals of great molecular volume. Characteristic minerals in the crystalline schists of this zone are orthoclase, all plagioclases, biotite, augite, olivine, garnet, cordierite, sillimanite, magnetite, and ilmenite. Many of the minerals of this zone also appear in the massive igneous rocks and in the contact-metamorphic rocks. The rocks are mostly gneisses, gradually approaching granites; also granulites, eclogites, and augite gneisses. Most of the igneous rocks are stable in this zone.

The orthoclase or microcline in the crystalline schists of the deepest zone tends to microperthite in the middle depths and to sericite in the upper zone. Plagioclases of the deep zone may be transformed into albite and anorthite and finally to albite and zoisite or sericite. The augites change to hornblende and finally

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[1] E. Riecke, Ueber das Gleichgewicht zwischen einem festen homogen deformierten Körper und einer flüssigen Phase, etc., *Nachr.*, Gesell. d. Wissensch., Göttingen, 1894, 4, pp. 278–284.

to chlorite. Olivine of the deep zone is transformed to hornblende or (with feldspar) to garnet and becomes serpentine in the upper zone.

Carbon dioxide and water doubtless escape from the deep zones upward wherever calcareous rocks containing free water or hydrated compounds become submerged in it.

Exceptional supplies of heat contributed by igneous intrusions may carry the reactions of the lower zones close to the surface.

Relation of Mineral Deposits to the Metamorphic Zones.—Though certain kinds of mineral deposits have originated at the surface or in the zone of weathering, the largest number have undoubtedly been formed in the zone of fracture, where circulation of solutions is comparatively easy. It is safe to assert that the great majority of ore deposits have been formed within 15,000 feet of the surface.

Ore deposits do not, as a rule, form in the zone of flowage or anamorphism where the passage of solutions is prevented. An exception to this is where hot emanations from intrusive bodies penetrate and impregnate certain rocks like limestone without the necessity of ducts and cavities.

Ore deposits may form also in the hottest zone where the solutions consist of magmas in which the free rearrangement of molecules is possible.

During the ordinary metamorphic processes under static or dynamic conditions extensive changes in mineral composition and structure may be effected with minimal changes in the chemical composition of the rocks, so that it is possible to trace the origin of highly metamorphosed rocks by the aid of analyses. Metamorphism can be, and usually is, effected with the aid of minute quantities of rock moisture and during the process there is little opportunity for extensive concentration of rarer constituents. Mineral deposits due to simple hydration or chemical rearrangement within the mass may result. Examples: soapstone by hydration of magnesian minerals; magnesite from carbonatization of serpentine; sulphur from reduction of gypsum by organic compounds; garnets developed in crystalline schists; concentration of hematite from lean primary ores; and many similar instances.

A comparison of the mineral records of ore deposits, formed at various levels in the earth's crust, with the results obtained by a study of general metamorphism soon brings out the fact that the same laws do not apply to both cases, although there are points of similarity. Attention was called to this important feature in a paper on the metasomatic processes in fissure veins, and increasing knowledge emphasizes the distinction.[1] Neither the rules of Van Hise nor the three zones of Grubenmann will fit closely the case of the ore deposits. The reason for this is not difficult to find. In metamorphism one deals with small quantities of solutions, free from large amounts of carbon dioxide and hydrogen sulphide. The majority of ore deposits, on the other hand, were formed by large quantities of waters rich in these gases and heavily charged with alkaline salts. A large number of silicates and other minerals, fairly stable under the influence of ordinary deep ground water, are incapable of existence in many vein-forming solutions. Biotite, amphibole, soda-lime feldspars, often also chlorite, serpentine, and magnetite are included among these.

Deposits Related to Igneous Activity.—Several important groups of ore deposits must therefore be considered apart from the ordinary processes of metamorphism. This especially applies to those deposits of the rarer metals which stand in closest genetic connection with intrusion and eruption of igneous rocks. It will be found that classified according to gangue minerals, they form three groups: (1) Those characterized by gangue minerals such as garnet, biotite, hornblende, pyroxene, specularite, magnetite, tourmaline, topaz, apatite, and scapolite, all associated with quartz; (2) those in which quartz, calcite, dolomite, siderite, barite, sericite, chlorite, and albite occur; (3) those in which quartz, chalcedony, opal, calcite, dolomite, barite, fluorite, sericite, chlorite, and adularia form the more abundant gangue minerals. Zeolites occur only exceptionally in these groups, and kaolin is considered almost wholly a product of descending waters.

The first group may be called the high-temperature deposits, though the highest temperatures during their genesis probably did not exceed 500° C. Ordinarily, but not necessarily, they were

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[1] W. Lindgren, *Trans.*, Am. Inst. Min. Eng., vol. 30, 1900, p. 601.

W. Lindgren, The relation of ore deposition to physical conditions, *Econ. Geol.*, vol. 2, 1907, pp. 105–127. Also Compte Rendu de laaX^{eme} session du Congrès Géologique international, Mexico, vol. 2, 1906, pp. 701–724.

formed at great depths and are sometimes designated as deep-seated deposits.

The second group is formed under intermediate conditions of temperature and in general also of depth.

The third group is formed at moderate temperatures, probably rarely exceeding 150° C., within a few thousand feet of the surface; many of them were developed close to the surface.

These groups are described in more detail in a later chapter.

On the whole it is evident that the great majority of ore deposits have been formed relatively near the surface and well within the zone of fracture, probably well within the upper 15,000 feet of the crust, and most of them within 10,000 feet of the surface; this applies even to contact-metamorphic deposits, many of which have developed along intrusive masses injected high into the zone of fracture. Some instances are known of such deposits having been formed within 3,000 feet of the surface. Only one class, that of the magmatic segregations, may have its origin at abyssal depths; but it is thought that more commonly the differentiation of these ores was effected after the intrusion of the magmas into the zone of fracture.

Derivation of Minerals.—Regarding the derivation of the valuable substances of mineral deposits we have to distinguish several groups:

1. Those which require little or no concentration, but merely involve chemical readjustment, as the formation of sulphur from sulphate.

2. Those which are formed by precipitation from solutions, the origin of which is beyond doubt, as, for instance, the salt beds derived from evaporation of sea water.

3. Those which are concentrated by mechanical means from well-known sources, like the gold placers.

4. Those which are derived by the solution and removal of waste material, like residual manganese or phosphate deposits.

5. Those which are derived from concentration in magmatic solutions by processes of differentiation; for example, certain titaniferous magnetites.

6. Those deposits, mainly of rarer metals, which require great concentration and concerning the origin of which more or less uncertainty still prevails.

Concentration.—Regarding the last group, it should first be stated that almost all rocks and inferentially all magmas contain small quantities of these rarer metals. There are two ways in which concentration of the rarer elements is possible. The first is by solution, by means of descending surface waters, of the small traces contained in the rocks and by corresponding deposition during the subsequent ascent of the waters. This solution of minor constituents is a slow and imperfect process, but that it actually occurs is shown by examination of natural waters and their deposits. The smaller the metal traces the slower and less complete is this process of solution. J. F. Kemp has studied this problem in some detail[1] and concludes that the amount which can be extracted by water percolating through the cracks of a rock is only one-sixth to one one-hundredth of the total amount of the particular metal contained. The leaching of compact masses of rocks by underground water is, therefore, at best an exceedingly imperfect process, and one on which it does not seem safe to rely for the concentration of the richer ores of the rarer metals like gold and silver. Veins and other deposits, conceded to have been laid down by purely meteoric waters, contain in fact little or no gold, silver, molybdenum, tungsten, and other rare metals.

Moreover, the descending waters are cool and dilute, and thus their chemical action is slow. To obtain effective concentration of such metals the first few thousand feet of percolated rock should probably be left out of consideration.

The second way of concentration is by disturbing the equilibrium of a molten magmatic solution. Such disturbances would take place by the irruption of magmas into higher levels of the crust or by cooling of the magma, or, in other words, by changes in pressure and temperature. From the study of volcanic phenomena it is known that under such circumstances certain substances are expelled from the magma, and that among these are water, halogens, alkaline salts, and a number of the rarer metals. From the study of plutonic phenomena we infer that a still more thorough expulsion of these substances was effected during the intrusions of deep-seated magmas. Gold, silver, lead, zinc, copper, molybdenum, tungsten, tin, and many other rarer metals certainly have a place in the list of magmatic emanations. Associated with these elements are the ions of sulphur, carbon, chlorine, fluorine, boron, and other elements.

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[1] J. F. Kemp, Problem of the metalliferous veins, *Econ. Geol.*, vol. 1, 1905–1906, pp. 207–232.

This concentration is effected automatically and with ease, and these elements, dissolved in water, ascend, propelled by the expansive force of the gases. The gaseous solutions will seek the fractures and fissures on their upward paths. Their high temperature facilitates the solution of other elements in the surrounding rocks. In upper and cooler levels the gases condense to liquid solution; precipitation begins by reduction of pressure and temperature or by reaction with the adjoining rock minerals. Finally, meteoric waters mingle with the magmatic and this again causes deposition and ultimately the still warm waters issue as ascending springs at the surface.

Both methods are used in the work of deposition by waters in the crust. The first is applicable to the more common gangue minerals and to the more abundant ore minerals. The second, it is thought, must be assigned as the main cause of the rich deposits of gold, silver, and other rare metals. The former class of minerals is found in all parts of the world; the latter is confined to districts where igneous forces have been active.

This conclusion is supported by an impressive array of observations of the various mineral deposits related to igneous rocks. The weight of the cumulative evidence is exceedingly strong, but is perhaps not fully appreciated, except by those who have made the study of these deposits their specialty.

UNDERGROUND TEMPERATURES[1]

The increment in temperature in the upper part of the earth's crust is generally assumed to average 1° C. for 30 meters or nearly 100 feet. Beginning with a surface temperature of 11° C. at a depth of 100 feet, corresponding to the mean annual temperature of a place in the temperate zone, we would have at a depth of 1,000 feet, 20° C.; at 9,000 feet, 100° C.; at 20,000 feet, 210° C.; and at 35,000 feet, 360° C., which is near the critical temperature of water (364° C.). As a matter of fact but little is known about the increment at great depths. Actual measurements within the accessible zone or to depths of about 6,000 feet show considerable divergences from the average figure given above. In some cases the increase in temperature is not quite uniform.

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[1] J. D. Everett, Evidence before the Royal Commission on coal supplies, London, 1904. Also in Reports of the British Association, 1882–1904. An increment of 1° C. in 100 feet equals 1° F. in 55 feet. The Royal Commission considered that the average would be 1° F. in 64 feet.

The best summary of the results obtained in widely separated parts of the world has been given by Koenigsberger,[1] who has also given important data regarding the influences which increase or diminish the geothermal gradient.

The results best available for general statements of the normal increment have, as a rule, been obtained from deep bore-holes in regions of slight relief, far from large bodies of water, and in little-altered rocks, with no Tertiary or post-Tertiary intrusions, and containing no large deposits of coal or oil. Observations in deep mines are probably somewhat vitiated by the cooling effect of ventilation; in new drifts and stopes, ventilation should not greatly affect the results obtained from bore-holes in the rocks.

The following data are abstracted from the tables of Koenigsberger:

GEOTHERMAL GRADIENTS IN UNALTERED ROCKS (NOT RECENT ERUPTIVES) AND REGIONS OF FLAT RELIEF. BORE-HOLES

Locality	Gradient in meters	Gradient in feet	Depth in meters	Depth in feet	Author
Martincourt,[2] France	31.0	101	1,200	3,937	
Sperenberg, Berlin	32.5	107	1,268	4,160	Dunker.
Sennewitz, Halle	36.6	120	1,048	3,438	
Schladebach, Merseburg	35.7	117	1,236	4,055	Dunker.
Paruschowitz,[2] Silesia	30.7	101	1,959	6,428	
Czuchow,[2] Silesia	29.6	97	2,239	7,346	Michael and Quitzew.
Bay City, Michigan	36.8	121	1,050	3,445	Lanc.
Marietta, West Virginia	37.9	124	1,360	4,462	Hallock.
Homewood, Pennsylvania	36.7	120	1,309	4,295	Cummins.
Wheeling, West Virginia	40.7	133	1,360	4,462	

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[1] J. Koenigsberger and M. Mühlberg, Ueber Messungen der geothermischen Tiefenstufe, *Neues Jahrbuch, Beil. Bd.* 31, 1911, pp. 107–157. (Contains also list of literature and technique of measuring temperatures.) Recent investigations relating to the measurement of temperatures in deep drill holes by maximum thermometers and thermo-electric methods are found in John Johnston and L. H. Adams, *Econ. Geol.*, vol. 11, 1916, pp. 741–762.

Note.—The deepest bore hole in the world is that of the Goff Farm near Clarksburg, West Virginia. On January 24, 1918, this had reached 7,350 feet and the temperature gradient is 1° F. in 51 feet. The boiling point of water should be reached at about 10,000 feet. U. S. Geol. Survey, *Press Bulletin* 357, 1918.

[2] At Paruschowitz, Czuchow, and Martincourt some coal beds are present.

The influence of cool bodies of water in lowering the earth temperature is shown in the following data:

Locality	Gradient in meters	Gradient in feet	Depth in meters	Depth in feet	Author
Port Jackson, N. S. W., Australia,	44.0	144	833	2,733	David.
Tokio, Japan...........	39.8	130	361	1,184	Tanakadate.
Pas de Calais, France....	56.6	185	1,400	4,593	Le Prince-Rinquet.
Copper mines of Lake Superior:					
a. Osceola, 8 km. from the lake,	42	138	303	994	Wheeler and Supan.
b. Atlantic, 3 to 5 km. from the lake,	52–55	171–180	276	905	
c. 1.5 km. from the lake,	67	220	508	1,667	
d. Close to the lake......	123	404	1,396	4,580	

Underneath high ridges and mountains the increase is slow:

Locality	Gradient in meters	Gradient in feet	Depth in meters	Depth in feet	Author
Mont Cenis (summit)....	50	164	....	....	Giordano.
Gotthard (summit)......	44	144	....	....	Stapff.
Simplon (summit).......	43.5	143	....	....	Schardt.

In or near recent eruptive rocks the increase is often rapid. This rapid increase is even noticeable in Tertiary eruptions. The following data are from borings:

Locality	Gradient in meters	Gradient in feet	Depth in meters	Depth in feet	Author
Sulz (Wurttemberg).....	24.1	79	710	2,329	Braun and Waitz.
Macholles, France......	14.2	46	1,005	3,329	Michel-Lévy.
Buda-Pest............	15.0	49	903	2,963	Szabó.

In the vicinity of heat-producing waters, or where chemical processes of decomposition are active, the increase is especially rapid.

Locality	Gradient in meters	Gradient in feet	Depth in meters	Depth in feet	Author
Idria, Austria	10	33	329	1,079	Scheinpflug and Holler.
Comstock, Nevada	17.1	56	457 672	1,499 2,205	G. F. Becker.

In coal mines and in borings in coal-bearing strata the increase is more rapid than the normal, owing to the chemical processes in the coal beds.

Locality	Gradient in meters	Gradient in feet	Depth in meters	Depth in feet	Author
Charmoy, Creusot (bore-hole).	26	85	1,168	3,832	Michel-Lévy.
Paruschowitz, Silesia (bore-hole):					
Above coal	26	85	1,122	3,681	Hendrick.
Below coal	35	115	1,959	6,427	Hendrick.
Gelsenkirchen, Germany	23.5	77	705	2,313	
Rosebridge, England (cooled by ventilation).	29.5	97	....	....	

Similar relations appear to exist in regions producing petroleum.[1]

Locality	Gradient in meters	Gradient in feet	Depth in meters	Depth in feet	Author
Santa Maria, California	23.0	76	1097	3,599	Arnold and Anderson.
Berekei, Caucasus	23.0	76	1000	3,281	Kelickij.
Apsheron, Russia	28.4	93	300–700	984-2,297	Solubjatnikow.

In mines of various kinds the increase may be more rapid or more slow than normal. The cooling by ventilation reduces the temperature to some extent.

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[1] For later information on this subject see H. Hoefer, Temperature in oil regions, *Econ. Geol.*, vol. 7, 1912, pp. 536–541 and E. De Golyer, The significance of certain Mexican oil temperatures, *Econ. Geol.*, vol. 13, 1918, pp. 275–301.

Locality	Gradient in meters	Gradient in feet	Depth in meters	Depth in feet	Author
Freiberg, Saxony............	31.0	102	457	ca. 1,500	D'Aubuisson.
Dolcoath, Cornwall..........	33.0	108	610	ca. 2,000	Haldane and Thomas.
Bendigo, Victoria (New Chum Railway).	42.7	140	1,110	3,645	Jenkins.
Ballarat, Victoria...........	44.2	145	634	2,080	Jenkins.
Witwatersrand, S. A.........	115	377	1,200	3,900	Marriott.

Although in some mines the increase is about normal, in other mines it is remarkably slow. At Bendigo,[1] where gold-bearing quartz veins occur in Ordovician sandstone, the rock temperature at the greatest depth attained, 4,600 feet, is only 112° F. (44.5° C.). At this depth the water is salty and has a temperature of 114° F. At St. John del Rey,[2] a gold mine in the schists of southern Brazil, at 4,000 feet below the adit tunnel, the temperature is only 95° F., or 35° C. On the Witwatersrand, in the Transvaal,[3] a temperature of 65° F. prevails at 500 feet; the increase down to 3,900 feet is regular at the rate of 1° C. per 360 feet; at the lowest depth the temperature is only 84.4° F., or about 29° C. According to Marriott the natural ventilation reduces the temperature near the workings 5° to 6° C.

No reason is known for the slow increase of temperature in the Transvaal and in Victoria. Koenigsberger has suggested that the decided increase in the earth temperature near oil pools and beds of coal (except anthracites) may be utilized for the prognostication of the occurrence of these substances near a given bore-hole.

In temperature measurements a maximum instrument constructed on the same principle as a clinical thermometer is most practical; an instrument about 25 centimeters long is recommended. An ordinary high-class chemical thermometer reading to 0.2° C. may also be used. For the measurement of tempera-

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[1] W. J. Rickard, Deep mining at Bendigo, *Mining Magazine*, London, 1910, pp. 281–282.

[2] *Eng. and Min. Jour.*, July 3, 1909.

[3] H. F. Marriott, An investigation of earth temperatures, etc., *Trans.*, Inst. Min. and Met., 1906.

See also *The Mining Journal* (London), April, 1906, p. 479.

ture in mine workings H. C. Jenkins[1] suggests some rules summarized below:

1. Temperatures should be taken in new workings which are rapidly pushed. Otherwise cooling may affect the result.

2. When possible the rock should be free from easily oxidized sulphides. Considerable heat is developed when pyrite oxidizes, as is well illustrated in many mines.

3. Holes should be bored 6 feet and, if possible, horizontal.

4. Wet ground should be avoided, as the readings will generally be too low.

5. One or two days should be allowed to permit the heat of drilling to dissipate.

6. The thermometer should be inserted, supported upon cork mounts in an outer closed glass tube, and the bore-hole closed.

Note 1.—In measuring rock temperatures in some Alpine tunnels[2] a hole 1.5 meters deep was bored in the side of the tunnel about a meter above the floor and slightly inclined upward. An ordinary thermometer was used, its length approximating 25 centimeters. It was cemented in a glass tube and inclosed in a double metal cylinder with cork rings. The mercury bulb was inclosed in a mixture of wax and turpentine. The metal cylinder with the thermometer was pushed in by means of a metal wire. The hole was then closed by a long plug of wood wrapped with woolen cloth, and stoppered by a wooden plug covered with gypsum plaster. The readings were taken 24 hours after the insertion. The errors or differences from the actual rock temperature are ±0.5° C.

Note 2.—A. C. Lane[3] doubts Koenigsberger's conclusion that the vicinity of Lake Superior affects the temperatures, and believes rather that climate changes may be responsible for the present slow increase. Lane gives the average gradient at the Calumet & Hecla as 189 feet for 1° C.

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[1] H. C. Jenkins, Rock temperatures in Victoria, *Proc.*, Aust. Assoc. Adv. Sci., vol. 9, 1902, pp. 309–318.

[2] E. Künzli, Geologische Beschreibung des Weissensteintunnels, *Beiträge zur Geologischen Karte der Schweiz*, Neue Folge, 21. Lieferung, Bern, 1908, p. 128.

[3] The Keweenaw series of Michigan, Lansing, 1911, p. 763.

CHAPTER VI

THE ORIGIN OF UNDERGROUND WATER AND ITS DISSOLVED SUBSTANCES

Origin of the Water.—There is no physical or chemical criterion by which the origin of a given water can be determined. A pure water might possibly rise from interior sources and acquire saline constituents during the ascent. A water of superficial derivation might be conceived to have become charged with magmatic products. If it is possible to distinguish between waters derived from the surface and those brought up from the interior of the earth, the evidence must be circumstantial and depend on geologic and physiographic testimony, such as geologic structure, igneous history, rainfall, and drainage basins. There are then two modes of derivation:

1. *Meteoric Waters.*—(*a*) The water is derived from the rain that falls on the surface, or from the water courses, or from the lakes, or from the present oceans and has simply descended into the earth in the cavities, fissures, or capillary openings to ascend at suitable places under hydrostatic conditions, or to remain stored in the rocks and almost stagnant (Chapter III). The term meteoric waters,[1] or surface waters, is applied to this group.

(*b*) The water was mechanically included in the sediments of ancient oceans and has for geologic periods been a constituent of these strata. The term "connate" has been proposed by A. C.

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[1] Reginald A. Daly, Genetic classification of underground volatile agents, *Econ. Geol.*, vol. 12, 1917, pp. 487–504. Daly shows that Posepny's term "vadose" (vadus = shallow) was applied by him to the descending waters above the water level, that is to the *zone of gathering* (Chapter III). Authors have used it since with different meanings, in each case including a certain part of the atmospheric waters. It seems as if the science could dispense with the word vadose. Daly following Archibald Geikie has suggested "epigene" to cover the underground activities of both fresh and marine waters. The term phreatic, applied by Daubrée to a somewhat indefinite part of the meteoric waters, may likewise be dispensed with.

Lane[1] to cover the origin of such waters, which really like those of the preceding class, are of meteoric origin.

2. *Magmatic Waters.*—The water existed in the solution constituting an igneous magma. Crystallization of the magma or its irruption into higher levels of the earth's crust liberated the water as one of the most volatile constituents, thus permitting its ascent to cooler levels. Such water may be called *magmatic* or *juvenile.*[2]

Underground waters may then be meteoric or magmatic or a mixture of both. Large quantities of magmatic water are rarely found except in regions of present or recent igneous activity.

Smaller parts of meteoric or magmatic waters may permanently or temporarily be withdrawn from the circulation by being held firmly by capillarity, by forming inclusions in minerals or by entering chemical compounds. Heat, pressure and chemical action may release part of these imprisoned waters when rocks sink into warmer zones or are engulfed by rising magmas. Thus while no one may doubt that the magma contains primary water a certain small part of it may be derived by the melting of rocks immersed in magmas.[3]

Magmatic or Juvenile Waters.—Volcanic phenomena are almost always accompanied by the emission of large quantities of steam and other volatile substances, and geologists generally have agreed that part of this water is a contribution to the atmosphere and hydrosphere from the magmas.[4]

More recently A. Brun[5] in a work of much merit on the volcanic exhalations has arrived at the result that the magmas are anhydrous, a view which is difficult to accept, though undoubtedly some classes of lavas like basalt, when arriving at the surface, are relatively poor in water. The clouds of vapors attending volcanic eruptions are, according to Brun, mainly volatilized chlorides, mixed with dust from explosions. Day and Shepherd[6]

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[1] *Bull.* Geol. Soc. Am., vol. 19, 1908, p. 502.

[2] E. Suess, Verh. der Gesell. deut. Naturf. und Aertze, 1902, pp. 133–150; Das Antlitz der Erde, Wien, Bd. 3, 2te Hälfte, 1909, pp. 630, 655.

[3] R. A. Daly, *Am. Jour. Sci.*, 4th ser., 1908, p. 48; Igneous rocks and their origin, New York, 1914, p. 249. Daly terms such waters *re-surgent.*

[4] T. C. Chamberlain, *Jour. Geology*, vol. 7, 1899, p. 559.

[5] Recherches sur l'exhalaison volcanique, Geneva, 1911. See also F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 282. A. N. Winchell, Brun's new data on volcanism, *Econ. Geol.*, vol. 7, 1912, pp. 1–14.

[6] Arthur L. Day and E. S. Shepherd, Water and volcanic activity, *Bull.*, Geol. Soc. Am., vol. 24, 1913, pp. 573–606.

recently disproved Brun's thesis by subjecting the gases of the Kilauea crater on the island of Hawaii to a very careful study, and ascertained that when free from contamination of air they consist of nitrogen, water gas, carbon dioxide, sulphur dioxide and hydrogen. They concluded that the water released from the liquid lava as it reaches the surface is entitled to be considered an original component of the lava with as much right as the sulphur or the carbon. It follows logically that some of this water from cooling lavas, with associated gases must mingle with the waters of meteoric origin.

Regarding plutonic rocks the direct evidence is lacking but indirect testimony is supplied by the inclusions of aqueous solutions so commonly found in granular rocks and by the presence of minerals like mica and amphibole which contain the hydroxyl molecule.

The best general evidence of the existence of juvenile waters is furnished, not by observation of the present springs, but by the study of old intrusive regions. Here the granites merge into pegmatite dikes, the latter change into pegmatite quartz, and this into veins carrying quartz and metallic ores, such as cassiterite and wolframite. Here we have evidence difficult to controvert that dikes consolidated from magmas gradually turn into deposits the structure and minerals of which testify to purely aqueous deposition; this admitted, it is difficult to see what would prevent such waters from reaching the surface in the form of ascending springs.

Elie de Beaumont[1] was the first to give full expression to this view. He believed that there were two classes of hot springs: The first (the more common) is intimately related to volcanism and derives its waters and dissolved solids from this source; the second, and more exceptional, derives its water from simple infiltration. This view was accepted by de Lapparent, but Daubrée arrived at the contrary conclusion, that both volcanism and thermal springs result from the infiltration of water from the surface; similar views were held by Fouqué and have more recently been adopted by de Launay.[2] The views of Daubrée found general acceptance in other countries; in the United States they were accepted by Le Conte, Van Hise, and others. All

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[1] *Bull.*, Soc. Géol. de France, Serie 2, 1847, Tome 4, p. 1272.

[2] L. de Launay, Recherche, captage et aménagement des sources thermominérales, Paris, 1892.

waters appearing at the surface were considered of atmospheric origin and their salts were dissolved from the rocks percolated. About the year 1900 the importance of magmatic exhalations for the formation of mineral deposits began to be reasserted by various mining geologists—among them Vogt in Norway, and Spurr, Kemp, Weed, and Lindgren in the United States. In 1902 Suess,[1] the eminent Austrian geologist, announced his belief that many of the springs in volcanic regions were of "juvenile" origin—that is, that they now reach the surface for the first time and yield a permanent addition of water and salts, carried up from magmas cooling at great depth. As an excellent example of this the Carlsbad Springs were cited.

The question now arises whether it be possible to establish criteria by which the magmatic waters may be distinguished from those of meteoric origin. Delkeskamp in Germany has attempted the solution of this problem in a series of suggestive papers.[2] He rightly considers temperature of little value as a criterion and points out that many springs of meteoric origin are hot, while some, strongly suspected to be of juvenile origin, are cold. The constant admixture with vadose waters forms another difficulty, but accounts well for the many derivatives of varying characteristics which accompany every spring of deep-seated origin. Seasonal variations of temperature, salinity, and quantity of water constitute excellent proofs of superficial origin. A practical constancy of salinity, temperature and quantity is said to be the best proof of a juvenile origin. Among the juvenile springs are those of Carlsbad in Austria, Ems and Wiesbaden in Germany.

It is doubtful whether these criteria can be accepted. Much more work must be done before we shall be able to establish the magmatic origin of any given spring.

Examples of Springs in Volcanic Regions.—As pointed out on p. 63 there are two types of ascending hot waters which may be

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[1] Verhandl. Gesell. deutscher Nat. u. Aerzte, Karlsbad, 1902.

[2] R. Delkeskamp, Juvenile und vadose Quellen, Balneologische Zeitung, 16, No. 5, Feb. 20, 1905, p. 15.

R. Delkeskamp, Die Genesis der Thermalquellen von Ems, Wiesbaden, und Kreutznach und deren Beziehungen zu den Erz—und Mineralgängen des Taunus und der Pfalz. Verhandlungen Gesell. deutscher Nat. und Aerzte, 1903, 2, First Part.

A. Gautier, *Compt. Rend.* vol. 150, 1910, p. 436.

See also reference in *Econ. Geol.* vol. 1, 1905, pp. 602–612.

of juvenile origin. They are the sodium carbonate and the sodium chloride-silica types, both common in regions of expiring volcanism. The former appear, for instance, in central Germany, in central France, in California and at various places in our Western States. The latter characterize the great geyser regions of Yellowstone Park, Iceland and New Zealand. The two classes break up through volcanic rocks and through the underlying plutonic rocks or crystalline schists. Whether these waters are wholly or partly of magmatic origin is a doubtful question. Arnold Hague, who spent many years in the study of the Yellowstone Park has expressed the decided opinion that the present hot springs at this locality are of meteoric origin.[1] Such an origin is probably more difficult to establish for the geyser district of New Zealand. On the other hand many geologists are of the opinion that some of the dissolved salts and gases at all of these places are of magmatic or juvenile origin.

Salts from Sedimentary Rocks.—There is little difficulty in the large sedimentary areas, where volcanism is absent. The great majority of underground waters are here simply of atmospheric origin, and, in spite of great diversity, the saline constituents, as well as the gases, are readily traced to the sediments traversed. It is evidently possible for atmospheric waters to attain sufficient depth to acquire a high temperature, though this rarely exceeds 60° C. The Hot Springs of Virginia, the Arkansas Hot Springs, the Arrowhead Springs of southern California, and the Utah Springs in the Salt Lake Basin clearly derived their saline constituents from the surrounding sedimentary rocks. Examples of this class from the French Alpine region are plentiful.

In all these waters the principal constituents are those of the surrounding sediments—calcium-magnesium carbonates from the limestones and dolomites, brines from the saline formations, calcium sulphate from gypsiferous Triassic formations, sodium sulphate from the Cretaceous shales, hydrogen sulphide from the reduction of sulphates by oil or other organic matter often present in the strata, carbon dioxide from reactions between carbonate of calcium and other salts. The presence of connate waters is difficult to prove. It is simply inferred from the occurrence of strong sodium chloride and calcium chloride brines in certain sedimentary rocks. Any marine beds must necessarily have con-

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[1] Origin of the thermal waters in the Yellowstone National Park, *Bull.*, Geol. Soc. Am., vol. 22, 1911, pp. 101–122.

tained occluded sea water, but many geologists doubt whether it would have remained undisturbed during long ages.

Salts from Igneous Rocks.—The various types of waters from sedimentary areas are closely paralleled by those from igneous rocks. The waters of the upper circulation in igneous rocks are characterized by their consistent content of calcium carbonate, to which large amounts of magnesium and ferrous carbonates are sometimes added. The attack of carbon dioxide on alkaline silicates gives alkaline carbonates and the oxidation of pyrite affords a small amount of sulphates. Soluble silica is added from all silicates, suffering partial decomposition. Occasionally these waters are tepid or hot, but probably only where they have exceptional opportunities for deep descent in regions of strong dislocations or contact with hot eruptive rocks.

Salts of Volcanic Springs.—Some of the hot ascending springs in volcanic regions carry much sodium carbonate as stated above. The long-continued action of the hot water saturated with carbon dioxide on the feldspars of the surrounding rock undoubtedly yields this salt in large quantities, and the scarcity of calcium and magnesium carbonates is explained by their precipitation with increasing percentage of alkaline carbonates.

Considerable quantities of sodium chloride are, however, always associated with the sodium carbonate and sometimes indeed predominate; to find an adequate explanation of this is more difficult. Igneous rocks average, according to Clarke's calculation, only 0.07 per cent. of chlorine, and while there are some exceptional rocks containing sodalite, the sodium chloride waters are by no means particularly associated with this mineral. Considering that the water could extract only a small part of this chlorine, it is not easy to estimate the amount of rock which must be percolated to obtain a sustained flow of chloride waters of the concentration often found in hot springs. The same reasoning applies to the alkaline sulphates which are abundant in some waters. It might be imagined that surface waters moving downward could have become charged with sodium chloride or sulphate while traversing saline sedimentary rocks, but such an explanation seems somewhat forced in the case of springs which issue from granite in a region where no such sedimentary beds are known to occur. Boron is a common constituent of many of these springs, for instance, the Steamboat

Springs, Nevada, and Ojo Caliente, New Mexico, both of which issue from granitic rocks. It is still more difficult to find a reasonable explanation for the presence of this substance on any hypothesis of leaching. Tourmaline and datolite are of course present in some rocks, but the springs carrying boron exhibit no marked relation to areas where such boron minerals occur. It is true that boron occurs in saline sedimentary beds and that traces of it are often found in waters traversing them, but the quantities do not compare with those determined in many waters of volcanic associations. Similar statements can be applied to fluorine, though it is less abundant than boron.

The geyser springs of Iceland, the Yellowstone Park, and New Zealand are rich in silica, and as most of them ascend through easily decomposed rhyolitic rocks, that substance may well be derived from leaching of the country rock. And yet when we note how veins rich in quartz are at places directly connected with pegmatite dikes, and how strong the evidence is against their deposition by leaching from surrounding rocks, we may well wonder whether this silica in the thermal waters is necessarily derived by solution of rock comparatively near the surface. And again, when we observe that chlorides form part of magmas, as indicated by the presence of sodalite, and remember that sodium chloride occurs as small crystals in the fluid inclusions of quartz phenocrysts, and finally note the abundance of chlorides at volcanic eruptions, would it not then be easier to account for this salt in the springs of volcanic regions by an easily effected concentration of volatile substances while the magma was still fluid, than by a laborious search for traces of chlorides in the congealed igneous rocks?

Origin of the Dissolved Gases.—Reduction of sulphates by organic matter and oil accounts satisfactorily for the hydrogen sulphide abundant in many sulphate and chloride waters in sedimentary rocks. Carbon dioxide occurs in many such waters, but the reactions by which this gas is produced are less well known. The rain water contains, of course, little carbon dioxide and far more is taken up during the percolation of the humus of the soil. This is soon absorbed in the formation of bicarbonates, but from the decomposition of these compounds gas may be set free.

In igneous rocks similar processes may apply on a small scale, but many of the ascending waters in volcanic regions contain such enormous amounts of carbon dioxide that its explanation by chemical reactions within the water itself appears utterly insufficient. From the earliest times of geological science this difficulty has been recognized and the literature on the derivation of the carbon dioxide is extensive. It has lately been summarized by R. Delkeskamp.[1] In addition to the carbon dioxide absorbed by the water from the atmosphere and the soil humus, a possible source of superficial origin may be found in decomposing organic deposits such as peat and lignite, though the coals are more likely to give off hydrocarbons than carbon dioxide. Still another way in which the latter gas may be formed is by the reaction between acid waters, such as are often found in mines, and limestone.

A more deep-seated source lies in the replacement near intrusive contacts of limestone or dolomite by calcic or magnesic silicates. This process has taken place, on a larger or smaller scale, at the contacts of all intrusives in calcareous rocks, and there is no good reason to doubt that it is going on, in some localities, at present. The quantity of carbon dioxide available from this reaction is large and it is likely, indeed, that many thermal springs in regions of intrusives have been fed from such sources. But, on the other hand, coal beds and limestones cannot be supposed to exist underneath every volcanic region, and in large areas of granite rocks the hypothesis becomes decidedly improbable.

It is known that many igneous rocks, particularly granites, contain liquid carbon dioxide as minute fluid inclusions, though it may be doubted whether they are as abundant as is assumed by some writers. On the assumption that the quartz contains a maximum of 5 per cent. by volume of these inclusions Laspeyres has calculated the content of CO_2 in a cubic kilometer of granite as sufficient to furnish the springs of Nauheim, Germany, with carbon dioxide for 273,000 years. Such calculations carry little conviction to those who realize the difficulty involved in the absorption of any but a minimal quantity of this gas from the quartz grains by percolating waters; and besides, exhalations of carbon dioxide are not characteristic of areas of granite, but appear in regions of volcanic rocks without reference to the character of the basement rock traversed (Fig. 4).

The interesting experiments of A. Gautier and others on the

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[1] R. Delkeskamp, Vadose und juvenile Kohlensäure, *Zeitschr. prakt. Geol.*, February, 1906.

gases included or occluded (absorbed) in the minerals of a rock and set free on heating have been summarized by several writers, including F. W. Clarke[1] and F. L. Ransome.[2] A great number of exact analyses of these gases were made recently by R. T. Chamberlin,[3] who found in general that the various pulverized rocks yielded a total amount of gases (at 0° C. and 760 mm. pressure) equal to from a fraction up to as much as 30 times the unit volume of rock; the gases determined were H_2S, CO_2, CO, CH_4, H_2, and N_2, among which CO_2 and H_2 always predominated. The carbon dioxide is believed to be derived from the decomposition of small quantities of secondary carbonates, while in part it may also be included or occluded. The hydrogen and other carbon compounds are probably due to reactions of water vapor and carbon dioxide with some of the substances

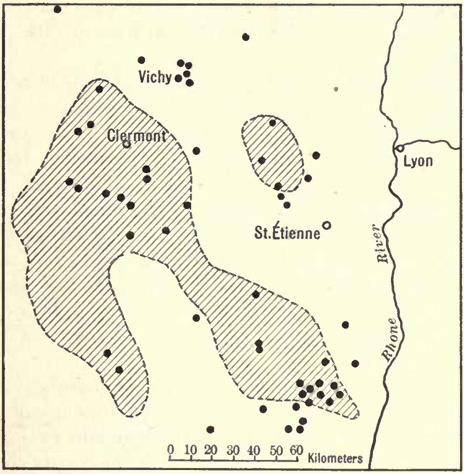

Fig. 4.—Carbon dioxide and sodium carbonate springs of central France. Black dots are springs. Shaded area shows extent of basaltic eruptions.

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[1] Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, pp. 276–281.

[2] *Econ. Geol.*, vol. 1, 1906, p. 688.

[3] R. T. Chamberlin, The gases in rocks, Carnegie Inst. Washington, 1908, p. 80.

contained in the rock, notably ferrous compounds. All of of these results are highly important and suggestive, and various hypotheses have been advanced showing how, by a sort of distillation, carbon dioxide and other gases might be given off and absorbed by ascending waters. There is indeed a possibility that some of the carbon dioxide in deep waters may have been derived in this way. But it is, perhaps, scarcely recognized that there is a great difference between heating a small quantity of pulverized rock in the air and obtaining the same amount of gases from a solid mass at great depth. It seems probable that pressure would prevent the escape of these gases, and if the mass of rock were heated to the melting-point it would undoubtedly acquire a capacity for absorption of far greater amounts of gases than those expelled by heating the powder to redness.

The presence of liquid carbon dioxide in cavities in minerals of igneous rocks is proof of its occurrence in the molten magma consolidated in depth. Every eruption brings new evidence of exhalations from magmas congealing near the surface; and almost every volcanic district of recently closed igneous activity testifies to the persistence of this gas in escaping from the cooling lavas below. The Cripple Creek district, where gold-tellurium veins cut through the core of an old volcano, presents an excellent illustration of this condition. Imperceptible at the surface, exhalations of carbon dioxide become more marked in depth and their temperature, higher than that of the surrounding rocks, indicates that they came from below. In the extinct volcanoes of the Auvergne in France and of the Eifel on the Rhine, waters highly charged with carbon dioxide and exhalations of the same gas are extremely abundant. It seems difficult to escape the conclusion that the enormous quantities of this gas contained in the ascending waters of volcanic regions are of igneous origin—a volatile constituent of the magma released when the magma was brought up to higher levels of less pressure in the earth's crust.

These considerations apply equally well to the hydrogen sulphide with which some of these springs are so abundantly supplied. The decomposition of sulphates by organic matter or other reducing agents may be appealed to in places, but in igneous rocks, like granite, it does not appear to be quantitatively sufficient, and as we know that this gas plays a prominent part in volcanic eruptions we may well feel justified in believing that the waters ascending in regions of such eruptions may absorb this gas or alkaline sulphides and carry them to the surface.

Rarer Elements Contained in Waters.—In the meteoric waters in crystalline rocks the salinity is low and rarer metals are generally absent. There are usually some iron and manganese, and possibly refined methods might discover other heavy metals in extremely small amounts. Where these waters appear in mines they naturally take up certain amounts of the metals of the deposits. Careful examination of calcium carbonate waters in sedimentary rocks has disclosed traces of nickel, cobalt, copper, lead, zinc, strontium, and barium, much more rarely fluorine, boron, and iodine. Arsenic is often present and where the waters precipitate limonite from dissolved ferrous carbonate, the ocher almost always contains traces of that metal. Where sulphates of iron and aluminum are among the saline constituents of waters in sedimentary rocks, determinable amounts of copper, zinc, cadmium, nickel, and cobalt may be found, with traces of lead, barium, and strontium. The quantity of cobalt usually exceeds that of nickel. Such waters issue at many places from beds of pyritic shale and owe their strong solvent power to the sulphuric acid generated by oxidation of pyrite.

Chloride waters from sedimentary rocks are always comparatively rich in bromine and barium, with traces at least of iodine and traces of boron, fluorine, and sometimes arsenic. From Wildbad, in Württemberg, in a strong sodium chloride water, traces of tin besides the substances mentioned, have been reported.[1] From the calcium chloride springs of Cannstatt, the same authorities mention nitric acid, boron, iodine, fluorine, barium, arsenic, and manganese; in the ochery deposits also copper, lead, and antimony or tin. From the cold sodium chloride water of Homburg, nickel, copper, arsenic, antimony, boron, and fluorine are reported; in the hot waters of Wiesbaden, copper, tin, arsenic, and boron. In almost all of the various waters mentioned above, traces of phosphoric acid are found.

In hot ascending sodium chloride springs which issue in volcanic regions rarer elements have often been determined; such waters are often rich in boron. Steamboat Springs, Nevada, contain notable amounts of arsenic and antimony with traces of quicksilver. The springs of the Yellowstone Park carry boron and arsenic, but are poor in other rarer constituents.

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[1] Stelzner and Bergeat, Die Erzlagerstätten, 2, 1905–06, p. 1220.

The ascending sodium carbonate springs in volcanic districts also frequently contain boron and fluorine in notable amounts. Arsenic and copper have been found in the springs of Ems, and the same metals with lead also at Vichy. The Carlsbad Sprudel contains, according to Göttl,[1] traces of bromine, iodine, fluorine, selenium, phosphorus, boron, barium, strontium, lithium, titanium, tin, arsenic, antimony, copper, chromium, zinc, cobalt, nickel, and gold.

The statements summarized above show clearly that traces of metals are by no means confined to springs of supposedly deep-seated, "magmatic" or "juvenile" origin, but that they occur in many different kinds of water. The presence of silver has apparently not been recorded, and that of gold only from the Carlsbad Springs. Quicksilver and large quantities of antimony seem to occur only in sodium chloride or sodium carbonate waters of the volcanic type, of which also higher amounts of boron and fluorine are characteristic. Arsenic is probably the most common of the rarer metals and has been found in all kinds of water. Copper, zinc, nickel, and cobalt are not uncommon, both in waters of sedimentary and in those of igneous origin. Lead is of rare occurrence. In minute quantities it is contained, together with zinc, in some calcium carbonate waters issuing from Paleozoic rocks of the Mississippi basin. Iron is characteristic of meteoric waters and occurs only in minute quantities in the hot ascending sodium springs.

The Igneous Emanations.[2]—At several places above the igneous emanations have been mentioned and the inference drawn that the waters in the crust of whatever origin must at places have absorbed such volatile substances. It may be well to describe briefly the character of these emanations.

The active volcanoes constantly emit volatile matter from lava flows, craters, and fumaroles. Some of the less volatile materials crystallize as sublimates near the gas vent; other parts escape into the atmosphere.

Chlorides[3] are given off in abundance by the erupted lavas

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[1] J. Roth, Allgemeine und chemische Geologie, Berlin, vol. 1, 1879, p. 570.

[2] For summary on this subject see F. C. Lincoln, Magmatic emanations, *Econ. Geol.*, vol. 2, 1907, pp. 258–274.

[3] According to Brun congealed lavas, upon heating, give off considerable quantities of chlorides, chlorine, hydrochloric acid, sulphur dioxide, and nitrogen. This statement still awaits confirmation.

and crystallize near the fumarolic vents; they comprise the salts of sodium, potassium, aluminum, ammonium, iron, copper, lead, and manganese. Sulphates come next in abundance; more rarely are fluorides or oxyfluorides found. Boric anhydride and sulphur are common products of sublimation at many places. Selenium and tellurium have been recognized in sulphur of volcanic origin. Arsenic in the form of realgar is reported, and the presence of cobalt, tin, bismuth, and molybdenum has been established, most of the detailed work having been done at the Italian volcanoes. Among the sulphides, pyrite, pyrrhotite, and galena[1] are mentioned. Specularite is a common product of the reactions effected by the volcanic gases. Even silicates, such as leucite, augite, hornblende, and sodalite, may be formed by sublimation.

Among the volcanic gases nitrogen, carbon dioxide, and hydrogen sulphide[2] are the most important and their emission, particularly that of carbon dioxide, may continue long after the igneous activity has subsided and even after the volcanic cone has been eroded. Evidence of this is furnished by the exhalations of carbon dioxide and nitrogen in the mines at Cripple Creek; of nitrogen at Creede and Tonopah; of carbon dioxide in the Tertiary gold deposits of New Zealand.

Some fumaroles in volcanic regions give off superheated steam with many associated salts and gases. Very interesting are the "soffioni"[3] of Toscana, Italy. These are vents emitting superheated steam of a temperature of up to 190° C. and a pressure of 2 to 4 atmospheres. They contain much boric acid undoubtedly of volcanic origin.

If all these substances are given off at the surface the intruding masses in depth must part with still more of their volatile constituents. All of these, whether from lavas or intrusives, must to greater or lesser degree mingle with the ground water. Beyond all doubt, ascending waters in regions of igneous activity must in places carry a considerable load of magmatic eamnations.

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[1] F. Zambonini, Mineralogia Vesuviana, Naples, 1910.

A. Bergeat, Die Aeolischen Inseln, Abh. math-phys. Klasse, K. bayer. Akad., 20, 1, 1899, p. 193.

[2] Oxygen is usually and marsh gas (CH_4) sometimes present.

[3] P. Toso, Bolletino del R. Comitate Geol., vol. 42, Rome, 1913, p. 122.

CHAPTER VII

THE SPRING DEPOSITS AT THE SURFACE

Processes of solution and precipitation are in continual progress in the underground waters. By a comparison between the products of the laboratory and those of nature we have arrived at the conclusion that a majority of mineral deposits have resulted from reactions in the underground water channels. Only at the surface, however, is it possible to study the actual progress of these chemical changes, and great interest, therefore, attaches to the deposits formed by the natural waters where they issue as springs from their underground path. The precipitation taking place in rivers, lakes and seas will be described in later chapters.

On the whole the composition of the material deposited by springs is simple. Three main divisions are recognized: Deposits of *limonite* (iron hydroxide), *calcium carbonate* and *silica*. Mixtures of two or all of these substances are frequently observed. The deposits are known as 1. Ochers, 2. Tufas, travertines or calcareous sinters, 3. Sinters or siliceous sinter.

The precipitation is in part due to cooling or escape of carbon dioxide but algæ and micro-organisms frequently aid by secreting silica jelly, calcium carbonate, colloidal ferric hydroxides or manganese oxide.[1]

Deposits of Limonite and Calcium Carbonate.—Limonite is frequently deposited by superficial meteoric waters which contain ferrous carbonate and ferrous sulphate. Many such ochers contain a little manganese and traces of arsenic, nickel and cobalt. Other waters also deposit some limonite so that many sinters and tufas are stained by this compound. Analyses of such ochers are quoted by F. W. Clarke.[2]

Calcium carbonate is probably the most common spring deposit, though the ordinary dilute surface waters rarely are

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[1] W. H. Weed, *Ninth Ann. Rept.*, U. S. Geol. Survey, 1889, pp. 613–676.
H. Molisch, Die Eisenbakterien, Jena, 1910.
Jour. Chem. Soc., vol. 92, pt. 2, 1907, p. 888, abstract.

[2] Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 205.

able to form important precipitates. Hot carbonated waters issuing from limestone often deposit large masses of such tufa, covering many acres with thick terraced beds. The Mammoth Hot Springs in the Yellowstone Park offers a beautiful example of such tufa. The precipitates are almost pure calcium carbonate with a little magnesium carbonate. In many of these springs the calcium carbonate is the least soluble constituent which remains after the others have been carried away. Thus, the sodium chloride springs of Glenwood, Colorado, yield a considerable deposit, and the sodium carbonate springs of Ojo Caliente, New Mexico, which are very poor in calcium, deposited at their former point of issue a porous tufa containing over 90 per cent. of calcium carbonate. This carbonate is no doubt deposited in crystalline form, though it is usually fine grained. Such deposits are not always calcite, for the presence of aragonite has been proved in many spring deposits, for instance those of Hammam Meskoutine, in Algeria, and of Carlsbad, in Bohemia.[1]

Deposits of Silica.—At hot springs containing much silica, this substance is abundantly precipitated because of evaporation, through mixture with other waters, or, according to W. H. Weed, by the action of certain hot-water algæ. The material is deposited as a colloid jelly which subsequently hardens to opaline or chalcedonic silica. Such sinters are formed by the hot springs of the Yellowstone Park and may contain up to 95 per cent. of silica. Sodium is often present as chloride or carbonate. The Steamboat Springs of Nevada[2] deposit a sinter of pure silica or mixtures of calcium carbonate and silica, the latter being present as chalcedony, or small crystals of quartz. (See Fig. 5.) This sinter contains weighable quantities of sulphides of mercury, lead, copper, arsenic, and antimony; the presence of gold and silver was also determined, and traces of manganese, zinc, cobalt, and nickel were found. Antimony sulphide, Sb_2S_3, is deposited as the amorphous "metastibnite" in quantities large enough to color the sinter red in places. In a shaft sunk into the gravel immediately adjoining the granite hill from which the springs issue, Lindgren[3] discovered delicate crystals of stibnite covering

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[1] H. Vater, *Zeitschr.* Kryst. u. Min., vol. 35, 1902, p. 149.

[2] G. F. Becker, The quicksilver deposits of the Pacific coast, *Mon.* 13, U. S. Geol. Survey, 1888, p. 341.

[3] W. Lindgren, *Trans.*, Am. Inst. Min. Eng., vol. 36, 1906, pp. 27–36.

the pebbles and associated with thin crusts of black opal and grains of pyrite or marcasite.

The sinter of the Yellowstone Park often contains arsenic, especially in the form of scorodite ($FeAsO_4.2H_2O$), and near one of the springs which was impregnated with pyrite Weed[1] noted rhyolite that contained traces of gold and silver. On the whole, however, the Yellowstone spring deposits are poor in the rarer metals. The same author, associated with Pirsson,[2] reports the occurrence of orpiment and realgar with native sulphur in a siliceous sinter from the Norris geyser basin. De Launay mentions a deposit containing orpiment at St. Nectaire, Puy-de-Dôme, France.

FIG. 5.—Section of chalcedonic spring deposits from Steamboat Springs, Nevada. White areas microcrystalline quartz. Magnified 29 diameters. Crossed nicols.

A calcareous sinter deposited by an ascending sodium carbonate spring in the Geyser mine, Silver Cliff, Colorado, on the

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[1] Mineral vein formation at Boulder Hot Springs, Montana, *Twenty-first Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1899–1900, pp. 233–255.

[2] Occurrence of sulphur, orpiment, and realgar in the Yellowstone National Park, *Am. Jour. Sci.*, 3d ser., vol. 42, 1891, pp. 401–405.

2000-foot level, yielded traces of lead, copper, zinc, nickel, and cobalt. At Hammam Meskoutine, in Algeria, a similar spring, according to Daubrée, deposits tufas and pisolitic sinters in which, in the concretions, shells of calcium carbonate alternate with shells of pyrite; strontianite is deposited by the same spring.

Quicksilver, gold, and silver have been recognized in the spring deposits of the geyser districts in New Zealand. From the Whakarewarewa hot springs at Roturoa (sodium chloride-silica type) sinters have lately[1] been analyzed which yielded nearly 5 ounces of silver and about $1 in gold per ton.

FIG. 6.—Section of chalcedonic spring deposits, from De Lamar, Idaho, showing vegetable remains. Magnified 35 diameters. Ordinary light.

At De Lamar, Idaho, Lindgren found in rhyolite spring deposits of flinty chalcedony, which included casts of vegetable remains and yielded traces of gold and silver.[2] (Fig. 6.)

S. Meunier[3] reported 0.5 per cent. of cassiterite in siliceous

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[1] J. M. Bell, *First Ann. Rept.* N. Z. Geol. Surv., 1907, p. 100.

[2] The gold and silver veins of Silver City, etc., *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1898, p. 187.

[3] *Compt. Rend.*, vol. 110, 1890, p. 1083.

sinter deposited by a hot spring at Selangor, in the Federated Malay States, but this statement has lately been challenged.[1]

It has been shown that springs, hot or cold, may deposit limonite in abundance, with arsenic, manganese, and traces of other metals; and it is likewise proved that the carbonate and silica sinters of hot springs, particularly those of the NaCl or Na_2CO_3 type, contain small quantities of the rarer metals, including gold, silver, copper, lead, zinc, antimony, arsenic, tin, and quicksilver. In very few instances has commercial ore been obtained from spring deposits at the surface. Quicksilver ores have been mined in New Zealand and ores of iron and manganese have been utilized in rare instances. The evidence that such waters have formed workable ore deposits is therefore strong but hardly conclusive; the remarkable poverty in metals of the deposits of the springs in the Yellowstone National Park, for instance, will to many seem an argument against the hydrothermal theory of the genesis of ore deposits.

Deposits of Other Gangue Minerals.[2]—Calcite, quartz, chalcedony, and opal are common products of deposition at the surface, but besides these the mineral deposits often contain such minerals as barite, ankerite and siderite, fluorite, and more rarely gypsum, strontianite, celestite, and zeolites. It will be necessary to examine the competency of the various waters to form these gangue minerals.

Fluorine is present in traces in many waters, both vadose and deep, but appears in larger quantities in waters of the sodium carbonate type. Few authenticated instances of actual deposition of fluorite by springs are recorded; the substance rarely occurs in crystallized form and the chemists have probably often neglected to test the sinters for fluorine. The Carlsbad Springs deposit a pisolitic sinter of aragonite and calcite. According to Berzelius[3] and later chemists this contains a notable quantity of calcium fluoride. A limonitic variety of the spring deposit yielded 0.272 per cent. arsenic.[4] The analyses on page 104 also demonstrate that various phosphates may be precipitated as well as the carbonates of iron and strontium.

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[1] J. B. Scrivenor, Origin of tin-deposits, Perak, Chamber of Mines, p. 5.

[2] The best résumé of the older data regarding spring deposits are found in Roth's Allgemeine und Chemische Geologie, vol. 1, 1879, pp. 564–596.

[3] *Pogg. Ann.*, 74, 1823, p. 149.

[4] Blum and Leddin, *Am. Chem. Pharm.*, vol. 73, 1850, p. 217.

COMPOSITION OF DEPOSITS OF CARLSBAD SPRINGS
Berzelius, Analyst

	A	*B*
Ferrous carbonate	12.13	....
Ferric oxide	19.35	0.43
Manganese oxide	....	trace
Calcium carbonate	53.20	96.47
Strontium carbonate	....	0.30
Basic ferric phosphate	1.77	....
Aluminum phosphate	0.60	0.10
Calcium phosphate	....	0.06
Calcium fluoride	....	0.99
Silica	3.95	....
Water	9.00	1.59

At Plombières, in the French Vosges, springs with a temperature of 70° C. issue from granite. They have a low salinity (360 parts per million) and contain mainly sodium sulphate and silica, believed to be present in part as sodium silicate, also traces of arsenic and fluorine. The derivation of these salts is doubtful and the springs are apparently not directly related to volcanic rocks. They issue from well-defined fissure veins containing quartz and fluorite, and Daubrée[1] found that the waters had actually deposited calcite, aragonite, and fluorite. The bricks and cements used by the Romans 2,000 years ago in the construction of the baths at Plombières were found to contain zeolites, chiefly apophyllite (containing fluorine) and chabazite, with opal and chalcedony. Chabazite is also reported by Daubrée as deposited from springs at Luxeuil and at Bourbonne-les-Bains, which have a temperature of 46° and 68° C. respectively.

In some of the mines of Cripple Creek, Colorado, gypsum is found suggesting deposition by ascending springs. Crystals of gypsum occur commonly near springs charged with calcium sulphate. Weed[2] has described how the Hunter Hot Springs, near Livingston, Montana, deposit this mineral in fractures in Tertiary sandstone; stilbite, a zeolite, is forming with the gypsum. The springs have a temperature of 64° C. and are weak mineral

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[1] Les eaux souterraines, 3, p. 31.

[2] Economic value of hot springs and hot springs deposits. *Bull.* 260, U. S. Geol. Survey, 1904, pp. 298–604.

waters. According to a somewhat doubtful analysis they are rich in silica and alumina, but poor in calcium sulphate, so that Weed believes that at present they deposit more stilbite than gypsum. The presence of stilbite has also been noted by Weed[1] in the vein-like deposits, containing gold and silver, believed to be made by the present Hot Springs at Boulder, Montana; the stilbite occurs in the predominating quartz, chalcedony and calcite. Lindgren noted the presence of a little adularia in the material.

According to Lindgren[2] a spring deposit in New Mexico contains about 89.60 per cent. of calcium carbonate and 0.9 per cent. of calcium fluoride. There are no springs now at this place, but it is probable that the sodium carbonate of Ojo Caliente, a short distance lower down in the valley, formerly issued here. As shown by an analysis on page 61, the water contains a notable amount of fluorine. A vein of white crystalline fluorite is opened by a shaft close by the calcareous tufa and is believed to have been formed by the same waters. Both tufa and vein contain traces of gold and silver, and a few crystals of barite were observed in the vein material. W. H. Emmons and E. S. Larsen[3] have described a similar case from Wagon Wheel Gap, Colorado.

Veins and replacements of fluorite in quartz porphyry and Cretaceous sandstone near the sodium carbonate springs of Teplitz, Bohemia, have been described by J. E. Hibsch[4] and the evidence is convincing that fluorite was deposited by these thermal waters.

Barite is deposited far more abundantly than fluorite. As shown above, many carbonate and even sulphate waters contain a little barium. It has been proved that alkaline bicarbonates with an excess of carbon dioxide can hold barium in solution, notwithstanding the presence of sulphates; sodium chloride and other salts also seem to retard the formation of barium sulphate. Haidinger observed that barite was deposited by the hot waters at Carlsbad,[5] and Becke noted the same at Teplitz.[6] At Idaho

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[1] Mineral vein formation at Boulder Hot Springs, Montana. *Twenty-first Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1899–1900, pp. 233, 255.

[2] *Econ. Geol.*, vol. 5, 1910, pp. 22–27.

[3] *Econ. Geol.*, vol. 8, 1913, pp. 235–246.

[4] *Tsch. M. u. P. Mitt*, 25, 1906, pp. 482–488.

[5] *Jahrb.* K. k. Reichsanstalt, 5, 1854, p. 142.

See also Delkeskamp, R., Entstehung und Wegführung des Baryts. *Notizblatt d. Ver. f. Erdkunde* (Darmstadt) (4), 21, 1900, pp. 55–83.

[6] *Tsch. M. u. P. Mitt.*, 5, 1883, p. 115.

Springs, in Colorado, a hot sodium carbonate water issues from granitic rocks, and barite crystals are found in abundance in a cellular and decomposed dike rock at the mouth of the spring. Spurr[1] shows that the barium is contained in this dike rock and believes that the barite resulted from the reaction of the water on the rock.

Barium is, however, far more commonly contained in sodium chloride waters, particularly in the brines of sedimentary strata. Many writers record the deposition of barite from such waters, and it is probable that wherever this mineral appears in large quantities in mineral deposits waters of this type have been active.

An excellent example is reported from a mine near Clausthal, Germany,[2] where a spring of strong brine, which undoubtedly derived its salts from sedimentary strata, was encountered at a depth of 1,200 feet; this brine contained in grams per liter 67.555 sodium chloride, 10.509 calcium chloride, 4.360 magnesium chloride, 0.350 potassium chloride, 0.314 barium chloride, and 0.854 strontium chloride. When it was mingled with the ordinary mine waters, which carried sulphates, abundant precipitation of barium and strontium sulphate took place in the pumps and elsewhere, so that within a few years the capacity of the pipes became much reduced by this deposit. This compact material contained 92.44 per cent. barium sulphate and 4.32 per cent. strontium sulphate. The reaction is believed to be retarded by the presence of sodium and magnesium chlorides.

According to P. Krusch,[3] barite is precipitated in the pumps at some Westphalian coal mines by a similar reaction between strong salt brine from the Triassic sandstones and potable water with sulphates, each ascending on separate faults and each deriving its contents of salts from sedimentary strata. Veins of barite with small amounts of galena, pyrite, etc., appear in the Carboniferous and in the Devonian. At a lower horizon quartz veins contain galena and zinc blende, both kinds of deposits having been made, according to Krusch, by these saline waters. Simi-

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[1] J. E. Spurr, *Prof. Paper* 63, U. S. Geol. Survey, 1908, p. 165.

[2] *Lattermann,* Die Lautenthaler Soolquelle und ihre Absätze. *Jahrb.* Preuss. geol. Landesanstalt, 1888, pp. 259–283. Ref. in Stelzner and Bergeat, Die Erzlagerstätten, II, 1905–06, p. 1218.

[3] *Monatsberichte* Deutsch. geol. Gesell., 1904, p. 36; Ref. in *Zeitschr. prakt. Geol.,* 12, 1904, p. 252.

lar deposits of barite in the pipes of the pumping apparatus have been described from English coal mines.[1]

An account by Headden[2] describes an interesting group of springs on the North Fork of the Gunnison River, Delta County, Colorado. They are cold, but contain free carbon dioxide and a little hydrogen sulphide and are essentially sodium chloride waters. At least one spring contains barium and all of them yield a little strontium. The Drinking Spring has a total salinity of about 1,656 parts per million. Small quantities of calcium, potassium, magnesium, barium (0.0132 gram per liter), strontium (0.0066 gram per liter), lithium, manganese, ammonia, iron, aluminum, also a trace of zinc, are present in the order stated; also sulphuric acid radicle (0.6254 gram per liter), silica, boron, and bromine, the latter three in very small amounts. The spring deposits a calcium carbonate sinter, which was found to contain 5.42 per cent. barite, but no gypsum or sulphur.

Ferrous carbonate, or siderite, is sometimes observed, as in the Carlsbad "Sprudelstein" and in deposits of limonite formed in bogs and peat. Deposits of magnesium minerals are rare. H. Leitmeier[3] describes a deposit of hydrous carbonate of magnesia from the springs of Lohitsch in Styria; many springs, especially those whose waters have traversed sedimentary beds, contain organic matter and are probably competent to deposit hydrocarbons.

The more common gangue minerals in certain classes of veins are thus deposited by spring waters, particularly by the warm sodium chloride and sodium carbonate springs. There are, of course, a great number of gangue minerals like tourmaline, garnet, feldspars, and similar silicates which cannot be expected to develop in water under the conditions of temperature and pressure prevailing at the surface.

Summary.—The deposits of ascending springs of undoubted meteoric origin contain opal, chalcedony, calcium carbonate, limonite, hydroxide of manganese, barite, siderite and pyrite. They often deposit sulphur by the oxidation of hydrogen sulphide. The ochery deposits very frequently yield small quantities of arsenic, copper, lead, zinc, nickel and cobalt.

The springs of the sodium carbonate and sodium chloride-

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[1] J. T. Dunn, *Chem. News*, vol. 35, 1877, p. 140.

[2] The Doughty Springs, etc., *Proc.*, Colo. Sci. Soc., 8, 1905, pp. 1–30.

[3] *Zeitschr. Krystall. u. Min.*, vol. 47, 1909, p. 104.

silica type in volcanic regions yield abundant deposits of opal, chalcedony, quartz, calcium carbonate, limonite, barite, siderite sometimes also pyrite. They also deposit fluorite which is rarely if ever found in the sinters of meteoric waters and yield smaller quantities of quicksilver, antimony, arsenic, lead, copper, zinc, tin, silver and gold. The rarer metals are thus more prominent and the waters are particularly characterized by a relative abundance of borates, arsenates and fluorides.

The list of recognizable minerals deposited by springs at the surface is as follows: Sulphur, quartz, opal, chalcedony, limonite, wad, calcite, aragonite, siderite, strontianite, barite, gypsum, celestite, fluorite, scorodite, pyrite, realgar, orpiment, cinnabar, stibnite, chabazite, apophyllite and stilbite.

CHAPTER VIII

RELATIONS OF MINERAL DEPOSITS TO MINERAL SPRINGS

The deposition of many valuable minerals can be directly observed in nature: limonite, for instance, from the evaporation of water containing iron, or from precipitation in bogs and lakes; sulphur by the decomposition of hydrogen sulphide dissolved in water; residual deposits of limonite, nickel silicates, and pyrolusite by the decomposition of rocks by meteoric waters; common salt and borax by the evaporation of lake waters. A large class of deposits, such as the deep-seated veins containing metals and ores developing near intrusive contacts, we can never hope to observe in nature in the process of formation.

Ascending mineral springs are not uncommonly observed in mineral deposits, particularly in those which follow fissures, but caution must be used in attributing a genetic rôle to these springs. If we find such a spring in a contact-metamorphic deposit or in a vein of deep-seated origin, as a cassiterite vein, it would be unlikely indeed that this spring had anything to do with the formation of the deposit, for it could scarcely be assumed that the circulation of underground waters could be maintained in the same path during the many vicissitudes of deep erosion, involving the laying bare of rocks once many thousand feet below the surface. The formation of ore deposits usually occupies comparatively short epochs, and the agencies to which they owe their origin are evanescent phenomena.

In a rather large class of veins, however, of which we know that they were formed near the surface and in recent geological times, we may look with more confidence for a maintenance of the originating solutions, but even here it is well to investigate carefully; the spring may simply be a water of the upper circulation which selected the fissure as a convenient path.

The case of Plombières has already been mentioned (p. 104) and there seems to be little reason to doubt that the quartz-fluorite veins at this place have been deposited by the same hot waters which still issue from the fissures. The Triassic sandstone, covering the granite in that vicinity, is in part replaced by silica and also contains fluorite and barite. Daubrée[1] cites the frequent occurrence, in the Triassic beds of the Central Plateau and the Vosges, of veins and extensive silicification similar to that at Plombières. Barite, fluorite, and sometimes galena accompany the quartz.

According to Jacquot and Willm,[2] the sodium chloride springs of Bourbon-l'Archambault, at the north end of the Central Plateau region, issue from a fracture in Triassic strata, which contains quartz with galena, barite, and fluorite. Dikes of micaceous porphyry (minette?) follow the fissures. The waters have a temperature of 53° C. and the total solids aggregate 3,186 parts per million, of which 1,770 are sodium chloride. Bromine, iodine, fluorine, arsenic, and copper are present, and the saline constituents are attributed to the Triassic and Permian strata. The spring deposits contain earthy carbonates and 0.07 per cent. copper oxide. The springs of Contrexéville, which issue from the Triassic and carry traces of fluorine and arsenic, have a temperature of 11° C., and the salts, among which calcium sulphate and calcium carbonate prevail, also bear marks of derivation from the sediments.

The springs of Lamalou, near Montpelier, southern France, have a temperature of 34°–47° C. and 1,500 parts per million of total solids; the alkaline carbonates prevail, but they also contain calcium and magnesium carbonates, suggesting an admixture of the meteoric waters. Traces of barium, arsenic, copper, lead, nickel, and cobalt were determined.[3] These springs are believed to be genetically connected with the eruption of a neighboring basalt area and stand in close relationship to veins containing pyrite, arsenopyrite, and chalcopyrite in a gangue of quartz and barite, the exploitation of which had to be stopped on account of the fear of tapping large volumes of water. Barite is believed to be deposited by the present waters.

The sodium carbonate springs at Ems,[4] according to Delkeskamp, issue from a fissure which forms the extension of a quartz vein and contains chalcopyrite. Basalt occurs in the same vicinity.

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[1] Les eaux souterraines aux époques anciennes, p. 151.

[2] Les eaux minérales de la France, Paris, 1894, p. 107.

[3] L. De Launay, Recherche, etc., des sources thermominérales, 1892.

[4] *Verhandl.* Gesell. deutscher Nat. u. Aerzte, 1903, 2, first part.

Sandberger and Delkeskamp state that the hot sodium chloride springs of Wiesbaden are closely connected with a quartz vein containing tetrahedrite; veins of barite and calcite are common, as are impregnations of barite; the latter are attributed to earlier (Tertiary) spring waters.

Close connection with ore-bearing veins is also, according to Delkeskamp, indicated by the sodium chloride springs of Kreutznach, which issue close to a number of veins containing calcite, barite, and fluorite with ores of copper and quicksilver. Here, also, Tertiary strata higher than the springs are impregnated with barite, suggesting a considerable age and a formerly higher point of issue of the springs. The saline constituents of the water are believed to be derived from sedimentary rocks.

The evidence presented by Delkeskamp does not suffice to establish direct connection of the springs with the ore deposition, but the widespread occurrence of barite, in close association with strong sodium chloride springs, is assuredly suggestive.

Mineral springs with a maximum temperature of 26° C. have been opened at several places in the mines of Freiberg, Saxony, and are described in some detail by Stelzner and Bergeat,[1] but there is little reason to believe that they are genetically connected with the deposits. The same authors describe weak sodium carbonate springs, which were encountered in the veins of Joachimsthal, in Bohemia; their highest temperature was 28° C. This is only about 10 miles from Carlsbad and at a lower level. Posepny has suggested that the waters may be derived from the same general source which supplies the springs at Carlsbad. Finally should be mentioned the hot waters which broke into the copper mine of Bocheggiano[2] at Massa Maritima, Tuscany, at a depth of over 1,000 feet, and which had a temperature of 40° C. They contained from 769 to 2,053 parts per million of total solids, mainly sulphates of calcium and magnesium, with a notable amount of boric acid. The waters, except for the boron, are of the type of ordinary mine waters and may be of meteoric origin with an admixture of boron from volcanic exhalations. Warm springs have been encountered in the mines of Cornwall; and one of them in a tin vein near Redruth is said to have contained much lithium, which is not surprising considering the general distribution of lithium-bearing muscovite in that

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[1] Die Erzlagerstätten, 2, 1905–06, p. 1227.

[2] B. Lotti and K. Ermisch, *Zeitschr. prakt. Geol.*, 1905, pp. 206–247.

region. It seems difficult to believe that these springs are the remains of the waters which deposited the veins, for the veins were formed at a great depth and under high pressure and temperature at a remote geological time.

In the Cordilleran region of the United States examples of mineral springs in mineral veins are not so common as might be expected from the coexistence of a late mineralization and present abundance of thermal waters. One reason for this lack lies probably in the great physiographic changes which in most parts of this region have taken place in relatively late times and which would tend to lower or divert the discharges of the springs. At Silver Cliff S. F. Emmons[1] found issuing from the 2,000-foot level of the Geyser mine a strong spring of sodium carbonate water with free carbon dioxide, yielding small quantities of copper, lead, and zinc; the temperature was 26.5° C. The shaft was sunk to a depth of 1,850 feet in rhyolite tuff; at this depth, at the contact between the tuff and pre-Cambrian gneiss, a vein was found containing galena, zinc blende, tetrahedrite, argentite, etc., in a gangue of calcite, barite, and quartz. The water deposited a calcium carbonate sinter with traces of lead, zinc, copper, nickel, and cobalt. In this instance it is possible that the ascending water may have had a genetic connection with the deposit.

At the Comstock lode[2] hot waters were encountered at an early date and have made exploitation difficult. It can scarcely be doubted that these waters stand in causal relation to the vein and it is certain that they now dissolve and precipitate gold and silver, as well as pyrite. The heat of the lode has been attributed to oxidation of pyrite or to kaolinization of the feldspars of the country rock, but Becker has shown that it is clearly due to the ascending waters. Reid[3] has examined the evaporated residue from water collected on the 2,250-foot level of the C. and C. shaft. He found 2.92 milligrams of silver and 0.298 milligrams of gold per ton of solution. This water, which has a temperature of 46° to 81° C., contains 965 parts per million of solids, mostly sulphates of calcium and sodium but including

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[1] The mines of Custer County, Colorado, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, p. 461.

[2] G. F. Becker, Geology of the Comstock lode, *Mon.* 3, U. S. Geol. Survey, 1882, p. 230.

[3] John A. Reid, The structure and genesis of the Comstock lode, *Bull.* 4, California Univ. Dept. Geology, 1905, pp. 177–191.

133 parts of silica. This water is assuredly not one of the pure types of ascending waters; its composition is in the main the same as that of the ordinary mine waters and it may be a mixture of meteoric mine waters with a very hot ascending water. Particularly convincing of the competency of the ascending "volcanic" springs to deposits gold and silver-bearing veins are the data given on p. 105 in relation to the Ojo Caliente springs of New Mexico and those of Wagon Wheel Gap in Colorado. To this is added the evidence of the gold and silver-bearing sinters of New Zealand (p. 102) and Steamboat Springs, Nevada (p. 100).

The widely cited occurrence at Sulphur Bank, in Lake County, California,[1] is considered to furnish good proof of deposition of cinnabar by hot sodium chloride waters, also heavily charged with boron (analysis on page 61). The springs issue through Quaternary basalt in which cinnabar was deposited with opal as crusts along crevices, sometimes as delicate crystals loosely attached to the walls, or as impregnations of the porous basalt; the pyrite or marcasite was mostly disseminated in the rock, but occurred also as crusts alternating with cinnabar and opal. Melville found traces of gold and copper in the marcasite. At the surface no cinnabar was observed, but sulphur, derived from the oxidation of H_2S, was present. A few feet below the surface the cinnabar appeared and continued down to about 300 feet, into the sandstones on which the basalt rested. No quicksilver was found in the water, but no one who has studied the occurrence has doubted that cinnabar, pyrite, and opal have been precipitated from the water which still ascends in these channels. The gases dissolved in the water consist mostly of carbon dioxide, with hydrogen sulphide, hydrocarbon, nitrogen, and some ammonia. The evidence gains in importance when it is realized that the mineral combination and general mode of occurrence cited are characteristic of the quicksilver deposits of the Coast Ranges. A number of other instances of deposition of cinnabar by ascending waters are given in Chapter XXIV.

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[1] G. F. Becker, Geology of the quicksilver deposits of the Pacific slope, *Mon.* 13, U. S. Geol. Survey, 1888, pp. 251–268.

Joseph Le Conte and W. B. Rising, The phenomena of metalliferous vein formation now in progress at Sulphur Bank, *Am. Jour. Sci.*, 3d ser., vol. 24, 1882, pp. 23–33.

F. Posepny, The genesis of ore deposits, 2d ed., Pub. by the Am. Inst. Min. Eng., 1902, pp. 32–36.

Summary.—There is then, convincing testimony that deposits of quicksilver, antimony, arsenic, gold and silver, may be formed close to the surface by hot ascending waters of the kind related to volcanic phenomena. It is probable, indeed, that the majority of fissure veins which contain notable amounts of gold and silver together with sulphides of the baser metals have been formed by these waters. Of this, more conclusive evidence is yielded by the many water deposited veins which so frequently, like a metallic aureole, surround the areas of igneous intrusive rocks.

On the other hand it is certain that warm and even cold waters of the meteoric circulation in non-volcanic regions are likewise competent to form mineral deposits of the baser metals containing oxides and carbonates of iron and manganese, and sulphides of copper, lead and zinc with very small quantities of gold and silver. There is even evidence that such waters may develop deposits of minerals of vanadium, and uranium with radium (Chapter XX, p. 399). The waters most competent to perform this work appear to be the calcium carbonate solutions and the chloride brines which at the same time contain carbon dioxide and hydrogen sulphides.[1]

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[1] C. E. Siebenthal, Zinc and lead deposits of the Joplin region, *Bull.* 606, U. S. Geol. Survey, 1915, p. 154.

CHAPTER IX

FOLDING AND FAULTING[1]

FOLDS

Sedimentary beds and ore deposits contained in them are often bent, corrugated, and folded in more or less complex manner. Extensive folding is usually effected by horizontal or "tangential" thrust, but minor bends and monoclines (Fig. 7) may originate by thrust in any direction. In extreme cases any fold or bend may pass over into a break or fault. In folding on a large scale it is necessary that the sedimentary complex have beds of sufficient strength (competent beds) to transmit the thrust and support the structures; if the complex is plastic it will be deformed by flowage and no regular folds will result.

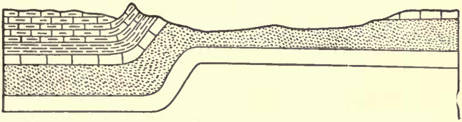

FIG. 7.—Monocline near Gallup, New Mexico. *After E. Howell.*

Folds are synclinal (Figs. 8 and 9), trough-like; or anticlinal (Fig. 10), shaped like a saddle. A plane which bisects the average angle between the limbs is called the axial plane of the fold. By complex movements the axial plane may become a curved surface. If this axial plane is vertical the limbs dip at like angles; if the axial plane is inclined the limbs have unequal dips. In close folding the limbs dip steeply (Fig. 11). When the axial plane of folds inclines strongly in one direction we speak

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[1] E. de Margerie and A. Heim, Les dislocations de l'écorce térrestre, Zürich, 1888, pp. 49–63.

Bailey Willis, The mechanics of Appalachian structure, *Thirteenth Ann. Rept.*, U. S. Geol. Survey, 1894, pp. 211–281.

C. R. Van Hise, Principles of North American pre-Cambrian geology, *Sixteenth Ann. Rept.*, U. S. Geol. Survey, pt. 1, 1896, pp. 589–633.

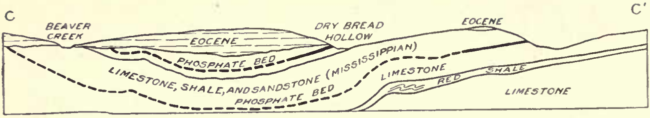

FIG. 8.—Open syncline showing Carboniferous phosphate beds unconformably covered by Eocene beds. Beaver Creek, Utah. *After E. Blackwelder, U. S. Geol. Survey.*

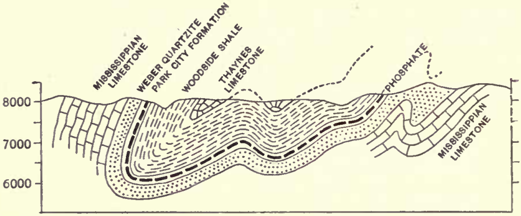

FIG. 9.—Eroded syncline, Georgetown Canyon, Idaho, showing phosphate bed. *After H. S. Gale, U. S. Geol. Survey.*

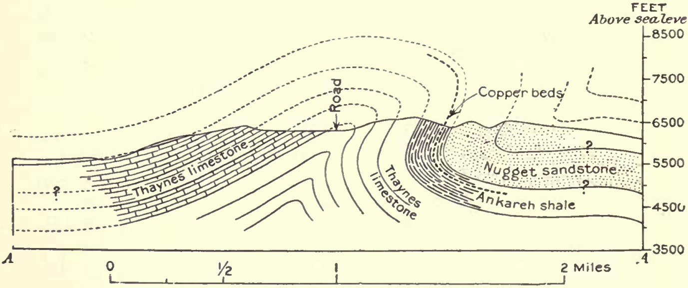

FIG. 10.—Eroded anticline, Montpelier, Idaho, showing bending of copper-bearing beds of Triassic age. *After H. S. Gale, U. S. Geol. Survey.*

of overturned folds, and these by further compression may easily pass over into overthrust faults (Fig. 12), causing a part of the folded series to slide over the other. In flat overthrust faults the horizontal movement may amount to many miles.

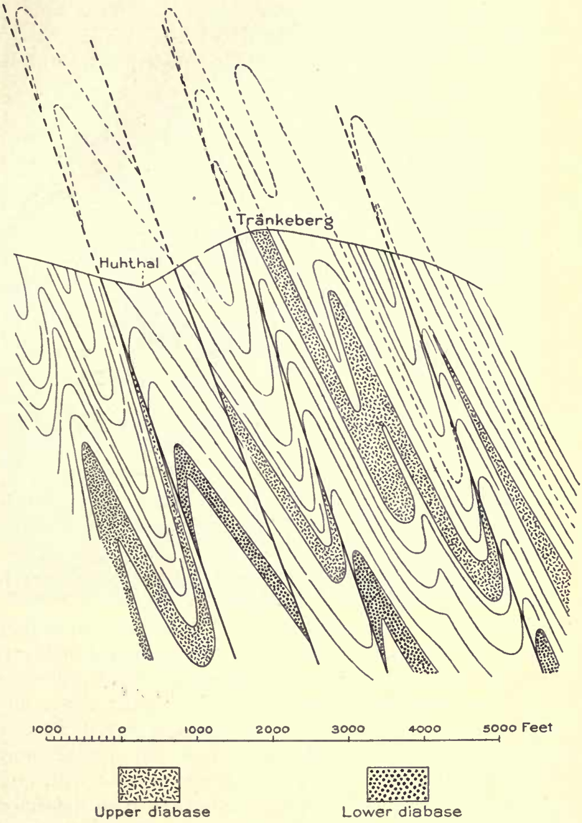

Fig. 11.—Close folding with overthrusts and thickening of strata by duplication. *After M. Koch.*

Synclines and anticlines extend naturally about perpendicularly to the direction of compressive stress. Their direction is indicated by a line passing through all the highest or lowest points of a given stratum. These crest lines or trough lines have usually a distinct dip; the angle of this line with the horizontal is called the *pitch* of the fold. Minor plications on the limbs often indicate the pitch of the fold. Thrusts in two directions result in cross-folding with, the development of bending in forms known as canoes, domes, and basins.

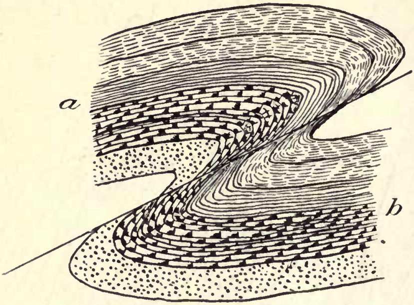

Fig. 12.—Diagram showing development of an overthrust fault from a fold. *After C. R. Van Hise, U. S. Geol. Survey.*

When beds are lifted in dome shape so that they dip away from a central point they form a *quaquaversal*.

In sharp folding of a sedimentary complex, the strata become thicker by compression at the points of greatest bending (Fig. 13). The harder strata of sandstone or limestone will yield to tension by breaking or tearing; the softer, shaly strata do not break, but yield to deformation. Strata of differing hardness may slide over one another at such points, and openings may be produced which, for instance, may later be filled with quartz. In slates and crystalline schists which have been deformed at great depths by flowage the harder or "competent" layers, like quartz veins, may be corrugated in an extraordinarily complicated manner. Some quartz veins of Nova Scotia, called "barrel quartz," are believed to owe their form to such conditions (Fig. 14).

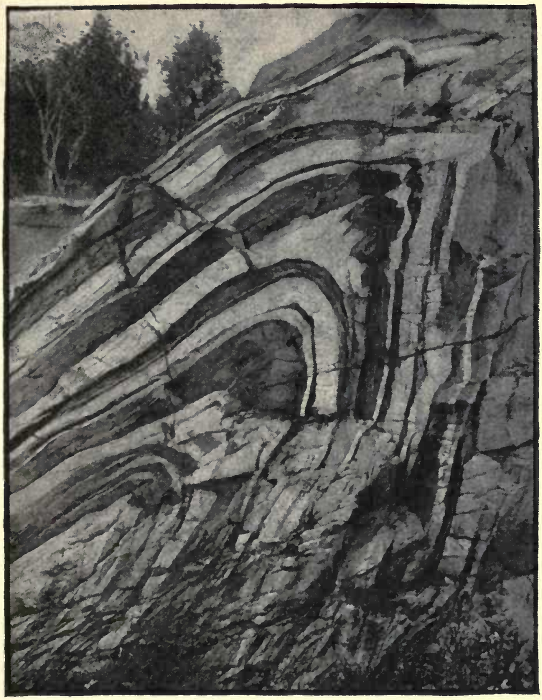

Fig. 13.—Overturned anticline of crystalline limestone, Lenox, Mass., showing thickening and breaking of strata at points of bending. *After T. Nelson Dale, U. S. Geol. Survey.*

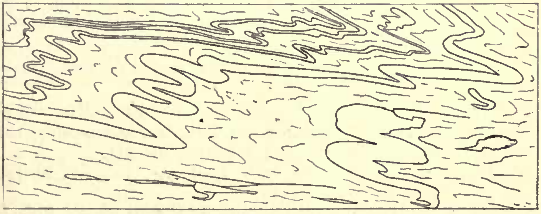

Fig. 14.—Gneiss with corrugated veinlets of quartz. *After C. R. Van Hise, U. S. Geol. Survey.*

FAULTS[1]

Sedimentary beds and deposits, as well as deposits of later origin which persistently follow a certain horizon in a sedimentary series, are sometimes abruptly cut off by structural planes.

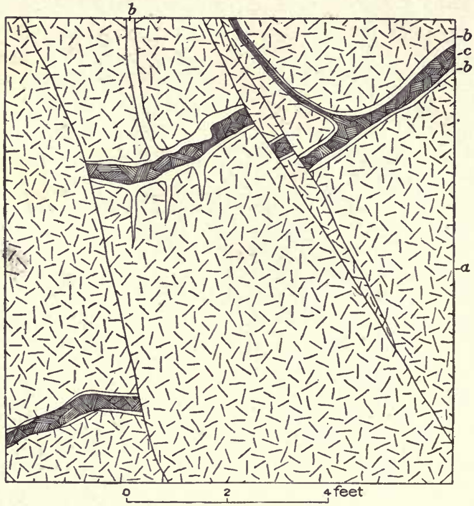

FIG. 15.—Faulting of Mendota vein, Silver Plume, Colorado. *a*, granite; *b*, quartz; *c*, galena and zinc blende. *After J. E. Spurr, U. S. Geol. Survey*

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[1] E. de Margerie and Heim, Bailey Willis, C. R. Van Hise, *op. cit.*

J. E. Spurr, Geology applied to mining, New York, 1904, pp. 149–163.

J. E. Spurr, Geology of the Tonopah mining district, Nevada, *Prof. Paper* 42, U. S. Geol. Survey, 1905, p. 144.

J. E. Spurr, Measurements of faults, *Jour. Geology*, vol. 5, 1897, p. 723.

J. E. Spurr, The relation of ore deposition to faulting. *Econ. Geol.*, vol. 11, 1916, pp. 601–622.

F. L. Ransome, The direction of movement and the nomenclature of faults, *Econ. Geol.*, vol. 1, 1906, p. 777.

C. F. Tolman, Jr., Graphic solution of fault problems, *Min. and Sci. Press*, June 17, 1911, *et seq.* Reprinted, San Francisco and London, 1911.

C. F. Tolman, Jr., *Econ. Geol.*, vol. 2, 1907, pp. 506–511.

H. F. Reid, Geometry of faults, *Bull.*, Geol. Soc. Am., vol. 20, 1909, pp. 171–196.

H. F. Reid, W. M. Davis, A. C. Lawson, and F. L. Ransome, Proposed nomenclature of faults, *Bull.*, Geol. Soc. Am., vol. 24, 1913.

C. K. Leith, Structural Geology, New York, 1913.

F. H. Lahee, Field Geology, New York, 1916, Chapters VII and VIII.

When such an occurrence is found it is safe to conclude that the interruption is due to a fault—that is, to a fracture along which movement has taken place—and that the continuation of the bed exists somewhere beyond this break (Fig. 15).

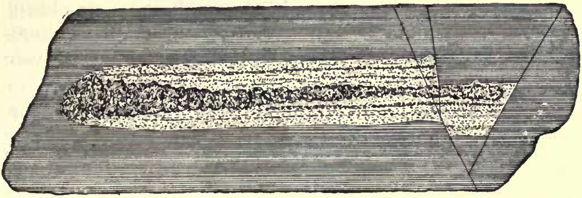

Fig. 16.—Sketch showing replacement of shale by pyrite. Natural size. The small fissures are older than the pyrite and are crossed by its banded structures. *After F. L. Ransome, U. S. Geol. Survey.*

In the case of epigenetic deposits not closely following the original lines of structure in the rocks, such a conclusion is often, but not always justified. The interruption of the ore-body may be due to an actual post-mineral dislocation, or it may be caused by a cessation of mineralization on account of structures existing before the mineralization began. The replacement of

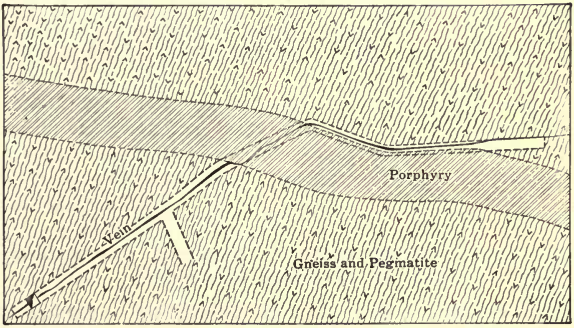

Fig. 17.—Plan of a vein in the Homer mine, Idaho Springs, Colorado, showing deflection of the vein upon meeting a dike. *After J. E. Spurr, U. S. Geol. Survey.*

limestone by galena or shale by pyrite (Fig. 16) may stop suddenly at a clay-coated seam, which offered a barrier to the solutions. A vein-filled fissure may terminate abruptly or split up within a few feet upon encountering softer and more plastic rocks, such as clay shales, thick gouge seams, or soft tuffs. A vein traversing formations of varying hardness often suffers abrupt deflection at rock contacts. It may also be deflected by encountering older dikes or fissures, either barren or filled with vein material (Fig. 17).

The distinction between faults and deviations is most important. The appearance of detached fragments of the ore—usually termed "drag"—on the faulting plane (Fig. 18), the direction of the striations, and the interrelations of dip and strike of faulting, fissure, and ore-body are all valuable data which must be interpreted in each case.

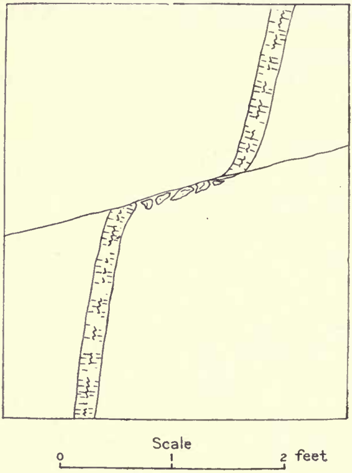

FIG. 18.—Sketch of faulted quartz vein in andesite, showing "drag." *After J. E. Spurr, U. S. Geol. Survey.*

There are many geometrical rules for the finding of a faulted ore-body, but they are of little value unless the character of the interruption is known. Each case must be considered and judged by itself.

Too often faulting is considered in only two dimensions—that is, as either normal or reverse movement in a vertical plane. The fact is that most faulting movements have lateral as well as vertical components; every mining engineer knows the frequent occurrence of inclined or horizontal striation on fault planes.

Several proposals have been made looking to a uniform nomenclature of the various elements involved in faulting movements; the best of these are advocated by J. E. Spurr, C. F. Tolman, Jr., and H. F. Reid. Lately a committee of the Geological Society of America has been instructed to examine this question in more detail, and their conclusions, in large part based on the work of Reid, will probably be adopted by American geologists. An abstract of their report[1] will be found in the following pages.

Measurements of fault movement are made in the fault plane itself, in a plane normal to the trace of the faulted body on the fault plane, in any normal plane, and in a horizontal plane.

General Terms

A fault is a fracture in the rock of the earth's crust, accompanied by a displacement of one side with respect to the other in a direction parallel with the fracture. The fracture is usually not an open crack, and an open crack would not be a fault unless one of the sides had moved parallel with the crack relatively to the other.

As we pass from one part of a fault to another, we find that certain characteristics vary. Definitions descriptive of characteristics must therefore be considered as referring to the parts of the fault to which they are applied and not necessarily to the fault as a whole.

A *closed fault* is one in which the two walls of a fault are in contact.

An *open fault* is one in which the two walls of a fault are separated. The same fault may be closed in one part and open in another.

The *fault space* is the space between the walls of an open fault.

A *fault surface* is the surface along which dislocation has taken place; this may be called a fault plane if it is without notable curvature.

A *fault line* is the intersection of a fault surface with the

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[1] Reid, Davis, Lawson, and Ransome, *op. cit.*

earth's surface or with any artificial surface of reference, such as a level of a mine.

When a fault is made up of a number of slips on closely spaced surfaces, the section of the earth's crust containing these minor faults is called the *shear zone.* This name would also apply to the *brecciated zone* which characterizes some faults.

The *fault breccia,* or *fault rock,* is the breccia which is frequently found in the shear zone, more especially in the case of thrust faults.

Gouge is a fine-grained, impervious clay, usually a mixture of minerals which sometimes occurs between the walls of a fault.

A horse is a mass of rock broken from one wall and caught between the walls of the fault.

The *fault strike* is the direction of the intersection of the fault surface, or the shear zone, with a horizontal plane. It is measured from the astronomic or from the magnetic meridian.

The *fault dip* is the vertical inclination of the fault surface, or shear zone, measured from horizontal plane.

The *hade* is the inclination of the fault surface, or shear zone, measured from the vertical; it is the complement of the dip.

The *hanging wall* is the upper wall of the fault.

The *foot wall* is the lower wall of the fault.

General Classification of Faults

Faults of *parallel displacement* are those in which all straight lines on opposite sides of the fault and outside of the dislocated zone, which were parallel before the displacement, are parallel afterward.

Rotatory faults are those in which some straight lines on opposite sides of the fault and outside of the dislocated zone, parallel before the displacements, are no longer parallel afterward—that is, where one side has suffered a rotation relative to the other.

Determinations of throw are almost always relative, and hence we can rarely tell which side of the fault has moved; therefore the terms "upthrow" and "downthrow," which are used according to the side from which the fault is viewed, are objectionable, as they suggest that a particular side of the fault has actually been moved. They are in very general use and should be retained, but it should be definitely understood that they refer merely to a relative and not to an absolute displacement.

Faults of Parallel Displacement.—No faults of any magnitude consist of simple parallel displacements over their whole length. Faults die out at their limits, and the displacement is not uniform along their courses, so that there is necessarily some slight rotation, varying in amount in the different parts of the fault's course. Probably the greatest number of faults, certainly of large faults, are of this character. The variations in rotation and displacement are permitted by slight plastic deformation. If, however, we confine our attention to a small length of the fault, we may describe the displacement there as if the rock were rigid; and if the rotation is very small, we may describe it as if a parallel displacement had occurred. It sometimes happens that the strikes on the opposite sides of a fault are different; the strata are then said to "strike at each other." This suggests a rotation, but it may be due to local variation of strike or to an unconformity.

The word "displacement" should receive no technical meaning, but is reserved for general use; it may be applied to a relative movement of the two sides of the fault, measured in any direction, when that direction is specified; for instance, the displacement of a stratum along a drift in a mine would be the distance between the two sections of the stratum measured along the drift. The word "dislocation" will also be most useful in a general sense.

There are two methods of defining the displacement due to a fault; we may define the apparent relative displacement of a bed by naming the distance between its two disrupted branches measured in any chosen direction, such as the vertical distance between the branches, measured in a shaft, or the perpendicular distance between the lines of intersection of the two branches with the fault plane; or we may define the actual relative displacement and its components in important directions. The apparent displacements are those usually measured; the actual displacement must be worked out for a complete understanding of the fault.

Only four important technical words are used to denote the various displacements caused by faulting, qualifying words being added to indicate the component of the displacement referred to. These words are:

Slip, which indicates the relative displacement of formerly adjacent points on opposite sides of the fault, measured in the fault surface. The qualifying words relate to the strike and dip of the fault surface.

Shift, which indicates the relative displacement of regions on opposite sides of the fault and outside of the dislocated zone. The qualifying words relate to the strike and dip of the fault surface, except in the expression "vertical shift," which is self-explanatory.

Throw, which indicates a displacement not related to the strike or dip of the fault plane.

Offset, which indicates the horizontal distance between the outcrops of a dislocated bed.

By keeping in mind the general meaning of these four words, all confusion in the uses of the proposed nomenclature can be avoided.

There is no generally accepted word in present use to denote the slip. Willis and Tolman use "displacement;" Spurr uses "throw." We have reserved "displacement" for general use, and the word "throw" is here used in quite a different sense. The word "shift" also suggests the meaning attached to it; there is no distinctive word now in use to describe the shift.

In mines, where the fault surface itself its visible, the slip will generally be determined; it is of paramount importance in mining. In field surveys, where the fault is studied by the dislocation of the outcrop of strata, or dikes, often at a considerable distance from the fault, the shift is determined. In the larger problems of geology the shift is of greater importance than the slip. The distinction between the slip and the shift is important; it has not heretofore been recognized in the nomenclature of faults. The perpendicular throw is of the greatest importance. It is frequently the only displacement determined, and in strike faults all the displacements in a vertical plane at right angles to the fault strike—that is, all the displacements which have heretofore received the most attention—can be expressed in terms of perpendicular throw. The offset is often the most important surface measurement made.

Faults in Stratified Rocks.—Among stratified rocks the character of the displacement of the strata due to a fault is so much influenced by the relation of the strike of the fault to the strike of the strata that syecial subclasses are generally recognized.

A *strike fault* is one whose strike is parallel to the strike of the strata.

A *dip fault* is one whose strike is approximately at right angles to the strike of the strata.

An *oblique fault* is one whose strike is oblique to the strike of the strata.

These terms are, of course, not directly applicable in regions of unstratified rocks; but they might be used in such regions with respect to the strike of a system of parallel dikes if this were distinctly stated in the description of the faults.

Similarly with regard to the general structure of the region:

A *longitudinal fault* is one whose strike is parallel with the general structure.

A *transverse fault* is one whose strike is transverse to the general structure.[1]

Slip.—The word "slip" indicates the displacement as measured on the fault surface; the qualifying words refer to the strike and dip of the fault.

The *slip* or *net slip* is the maximum relative displacement of the walls of the fault, measured on the fault surface, along the line of the movement; it is given by *ab* in Figs. 19 and 20.[2]

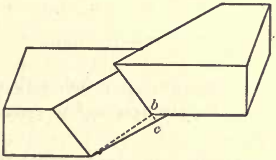

FIG. 19.—The slip.

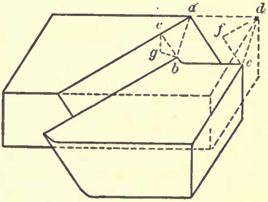

FIG. 20.—The shift.

The *strike-slip* is the component of the slip parallel with the fault strike or the projection of the net slip on a horizontal line in the fault surface; *ac* in Figs. 19 and 20.[3]

The *dip-slip* is the component of the slip parallel with the fault dip, or the projection of the slip on a line on the fault surface perpendicular to the fault strike; *bc* in Figs. 19 and 20.[4]

Shift.—It frequently happens that a fault has not a single surface of shear, but consists of a series of small slips on closely

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[1] See the word "flaw" further on.

Spurr and Tolman call this the "total displacement."

[2] Tolman calls this the "horizontal displacement."

[3] Tolman calls it the "normal displacement."

spaced surfaces, and in some faults the strata in the neighborhood of the fault surface are bent, so that the relative displacements of the rock masses on opposite sides of the fault may be quite different from the slip and not even parallel with it. The word "shift" is used to denote the relative displacements of the rock masses situated outside of the zone of dislocation; the qualifying words relate to the strike and dip of the fault, with one exception, in which the meaning is clear.

The *shift*, or *net shift*, is the maximum relative displacement of points on opposite sides of the fault and far enough from it to be outside of the dislocated zone; *de* in Figs. 20 and 21, where *d* is the position of a selected point before and *e* after the faulting.

The *strike-shift* is the component of the shift parallel with the fault strike; *df* in Figs. 20 and 21.

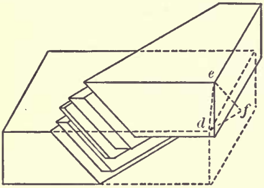

Fig. 21.—The shift.

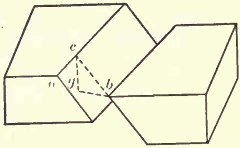

Fig. 22.—The throw.

The *dip-shift* is the component of the shift parallel with the fault dip; *fe* in Figs. 20 and 21. (The triangle *def* is parallel with the fault surface in Fig. 20.[1])

The bending of the strata near the fault may be so great that the direction of the shift is no longer even nearly parallel with the fault surface; it is better then to use the three following terms for the components of the shift:

The *strike-shift* is the horizontal component of the shift parallel with the fault strike, as already defined.

The *normal shift* is the horizontal component of the shift at right angles to the fault strike. It equals the horizontal shortening or lengthening of the earth's surface at right angles to the fault strike, due to the fault.

The *vertical shift* is the vertical component of the shift. These

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[1] The dip-shift and strike-shift are not accurately shown in Fig. 20, because the net shift, *de*, is not parallel with the fault plane, and the lines *de*, *df*, and *fe* would not lie in one plane. But the definitions are clear and the figure illustrates them fairly well.

terms may evidently be used equally well when the shift is parallel with the fault plane.

Throw.—The word "throw" will apply to components of the displacement having no immediate bearing on the strike or dip of the fault plane.

The *throw* is the vertical component of the slip; *cg* in Figs. 20 and 22, *de* in Figs. 23 and 24. The word is almost universally used in this sense, but A. Geikie uses it to designate the vertical distance between the two parts of a dislocated bed, projected if necessary—a very different thing. Geikie's "throw" would be represented by *df* in Figs. 23 and 24. Spurr uses "throw" to designate the distance between the two parts of a dislocated bed measured on the fault plane.

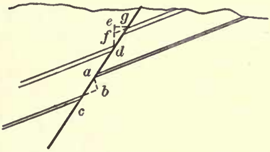

FIG. 23.—Section of a normal fault.

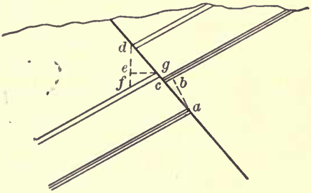

FIG. 24.—Section of a reverse fault.

The *heave*[1] is the horizontal component of the slip, measured at right angles to the strike of the fault; *bg* in Figs. 20 and 22, *eg* in Figs. 23 and 24. The word "heave" has been used in many senses; J. Geikie, Willis, Scott, and Fairchild use it as defined above; A. Geikie and Spurr use it to designate what we have called the "offset" of a bed (see below); Jukes-Brown apparently used it for the strike-slip (De Margerie and Heim, page 72); so did Ransome; and Scott also uses it in this sense when he refers to "heave faults."

The *perpendicular throw* of a bed, dike, vein, or of any recognizable surface, is the distance between the two parts of the disrupted bed, etc., measured perpendicularly to the bedding plane or to the plane of the surface in question. It is measured, therefore, in a vertical plane at right angles to the strike of the disrupted surface.[2] The importance of the perpendicular throw

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[1] Sometimes called the "horizontal throw."

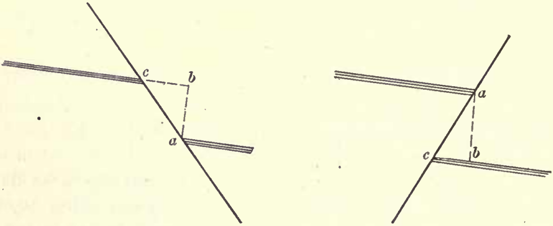
[2] Spurr calls it the "perpendicular separation." Tolman's "perpendicular throw" would under certain conditions correspond in meaning with our expression.

of the strata is so great that it is convenient to have special terms for it; these are given below.

The *stratigraphic throw* is the distance between the two parts of a disrupted stratum measured at right angles to the plane of the stratum; *ab* in Figs. 23 and 24. The stratigraphic throw is in general the simplest throw to determine; it can be found from the distance between the outcrops of the two parts of the same stratum, the dip of the stratum, and the slope of the ground.

The *dip throw* is the component of the slip measured parallel with the dip of the strata; *cb* in Figs. 23 and 24.

The throws have been defined as components of the slip. Where we are dealing with a simple fault in plane strata, the shifts will be the same as the slips, and the term throws will apply to both equally well; it is only in plane strata that the perpendicular throw is important. Where there is a dislocated zone about the fault, the term perpendicular throw would necessarily apply to the shift; but we cannot detach the word throw from its accepted meaning and apply it generally to the shift.



FIGS. 25 and 26.—Plan of an oblique slip.

Offset.—The *offset* of a stratum is the distance between the two parts of the disrupted stratum measured at right angles to the strike of the stratum and on a horizontal plane.[1]

If Figs. 25 and 26 represent the ground plans of oblique faults on a level surface, *ab*, and not *ac*, would be the offset of the stratum; *ac* would be the horizontal displacement of the stratum parallel with the fault strike.

Some confusion of nomenclature results from the non-observance of the fact that the distance between the dislocated parts of a stratum measured in a certain direction is not the same as

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[1] A. Geikie and Spurr use the term "heave" for this offset.

the component of the slip in the same direction; for instance, let Fig. 27 represent a reverse strike dip-slip fault[1] in section. A. Geikie calls *ad* the throw and *ef* the heave, whereas the most general usage seems to be to call *ac* the throw and *bc* the heave, as adopted above. The distance *ad* has not been defined, but it is readily described as the vertical displacement of the stratum, without limiting the word "displacement" to a technical meaning; *ef* is the offset. Let Figs. 28 and 29 lie in the fault plane and let the point *a* move by faulting to *c*, then *ac* will be the net slip, *ad* the strike-slip, *cd* the dip-slip, and *ab* the displacement of the stratum parallel with the fault strike; *ab* is not necessarily at right angles to the strike of the strata.

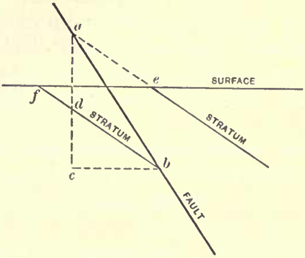
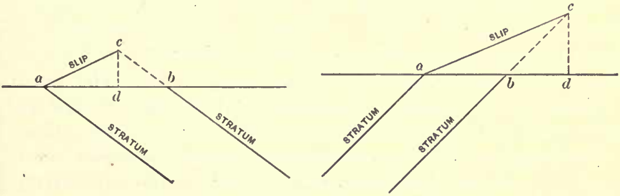


FIG. 27.—Plan of a reverse fault.



FIGS. 28 and 29.—Section in a fault plane.

Faults Classified According to the Direction of the Movement.—Faults may be classified, according to the direction of the movement on the fault plane, into three groups, as follows:

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[1] This means a reverse fault whose strike corresponds with the strike of the strata and in which the displacement has been in the direction of the fault dip.

Dip-slip faults, where the net slip is practically in the line of the fault dip.

Strike-slip faults, where the net slip is practically in the direction of the fault strike. J. Geikie calls such faults "transcurrent faults," or "transverse thrusts." Jukes-Brown calls them "heaves." Scott calls them "horizontal faults," or "heave faults." A vertical fault is one with a dip of 90 degrees (see below); and, by analogy, a horizontal fault should be one with a zero dip and the term should not be applied to strike-slip faults.

Oblique-slip faults where the net slip lies between these directions.

Classes of Strike Faults.—Most geological text-books and books on field methods have confined themselves almost exclusively to the discussion of dip-slip faults, and have given little attention to the other two classes.

Strike faults have usually been treated as if they were also dip-slip faults and classified into

Normal faults, where the hanging wall has been depressed relatively to the foot wall.

Reverse faults, where the hanging wall has been raised relatively to the foot wall.

Vertical faults, where the dip is 90 degrees.

The relative displacement has usually been determined by means of a dislocated bed. Although exception may well be taken to these terms, their retention is recommended, because they are in general use and are well understood. The word "reverse" is preferable to "reversed" (which has been almost universally used), as the latter implies the reversal of an action.

The horizontal distance between two points on opposite sides of a fault, measured on a line at right angles to the fault strike, is always shortened by a reverse strike fault, lengthened by a normal strike fault, and unchanged in length by a vertical fault. It can be shown that normal faults may be formed without the existence of tension and indeed under some pressure, but the definitions we are giving are geometric and not dynamic.

Extension of the Words Normal and Reverse to Diagonal and Dip Faults.—The expressions "normal" and "reverse" may be used in connection with oblique and dip faults, even when these are strike-slip or oblique-slip faults, provided they are applied to designate the apparent relative displacement of the two parts of a dislocated stratum, or other recognized surface, in a vertical plane at right angles to the fault strike. It does not follow, in the case of oblique-slip faults, that a horizontal line at right angles to the fault strike would be lengthened by a normal or shortened by a reverse fault. This has been pointed out by Ransome[1] and can be illustrated by Figs. 30 and 31. In Fig. 30 a reverse fault, as determined by the displacement of the stratum, has caused an extension at right angles to the fault strike. It is evident that if the hanging wall had moved, as in Fig. 31, with the stratum dipping as there represented, we should have had a normal fault and a contraction at right angles to the fault strike. The relations of the two parts of the disrupted stratum in Fig. 30 are exactly the same as if we had had a simple reverse dip-slip fault, and in Fig. 31 as if we had had a simple normal dip-slip fault; and if there are no disrupted dikes or other means of determining the amount of the strike-slip, the movements described could not be distinguished from simple dip-slip faults.[2] It very frequently happens that nothing more than the apparent displacement of the strata can be determined, and we recommend that the terms "normal" and "reverse" faults as defined be used purely for purposes of description and not for the purpose of indicating extension or contraction, tension or compression, vertical or horizontal forces.

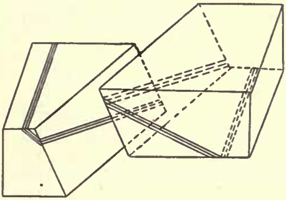

FIG. 30.—A reverse fault due to an oblique slip.

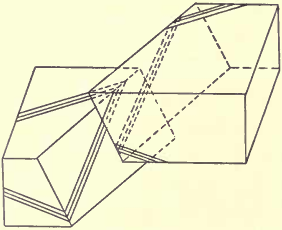

FIG. 31.—A normal fault due to an oblique slip.

Special Classes of Faults.—There are two kinds of faults which have played such important rôles in altering the structure of some regions that they have received special names.

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[1] *Econ. Geol.* vol. 1, 1906, pp. 783–787.

[2] The methods of determining the complete displacement at a fault are given in Reid's Geometry of faults, *Bull.* Geol. Soc. Am., vol. 20, 1909, pp. 170–196, and in Tolman's Graphic solution of fault problems, *op. cit.*

Overthrusts.—These are reverse faults with low dip or large hade. In some cases the dip-slip has been enormous, amounting to tens of kilometers. Scott calls them "thrusts" and separates them entirely from faults of high dip; but the word "thrust" has been used for ordinary reverse faults of high dip. The word "overthrust" has been generally used for this kind of fault and is very descriptive. It should be adopted.

Flaws.—Suess has described with care certain faults in which the strike is transverse to the strike of the rocks, the dip high and varying from one side to the other in the course of the fault, and the relative movement practically horizontal and parallel with the strike of the fault. He has used the word "Blatt" (plural, "Blätter") to designate them. Miss Sollas has used the word "flaw" in the English translation of Suess. The gold-quartz veins of the Tauern in Austria, investigated by Posepny, follow such dislocations.

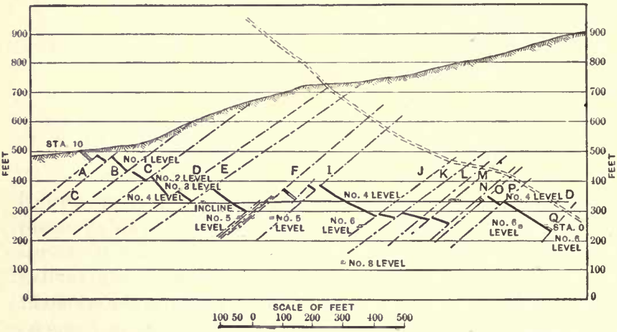

Fig. 32.—Vertical section of a faulted vein, Berlin mine, Nevada, showing also its probable original position. *After Ellsworth Daggett.*

Rotatory Faults.—When a rotation of one side of the fault occurs, the amount of the rotation and the direction of the axis should be given. Rotatory faults have been but little studied, and it is not considered advisable to suggest a special nomenclature at present.

Mineralization of Faults.—Any fault may become a *fissure vein* by filling and replacement along its course. However, it is rather unusual to find large structural faults, normal or overthrusts, which have been extensively mineralized. Shear zones, sheeted zones, and "flaws" (Blätter) often result in veins or lodes.

Complexity of Faulting.—During mining operations excellent and detailed instances of the complexity of faulting are often found. Normal and reverse faults may occur in close proximity. A fault consists more frequently of a series of closely spaced breaks than of a single fracture. Displacement occurs usually along each of these breaks, the result being a distortion of the deposit within the faulted zone.

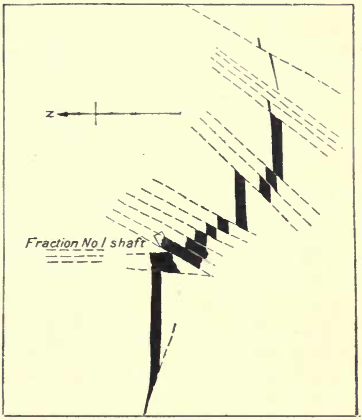

Fig. 33.—Horizontal plan showing faulted vein, Tonopah, Nevada. Scale 50 feet to one inch. *After J. E. Spurr, U. S. Geol. Survey.*

Fig. 32 shows a case of complicated normal faulting from the Berlin vein, Nevada.[1] Besides the faults indicated there are a great number of other dislocations with horizontal displacement. The deposit is a filled quartz vein, 2 to 3 feet wide, carrying 2 per cent. of sulphides with silver and gold.

The great complications ensuing where faulting takes place along two intersecting fault systems have been described by Spurr[2] in his report on the Tonopah district, Nevada. The

[1] Ellsworth Daggett, The extraordinary faulting at the Berlin mine, *Eng. and Min. Jour.*, Mar. 30, 1907.

[2] Prof. Paper, 42, U. S. Geol. Survey, 1905.

result of such structures is likely to be a zigzag distribution of the fragments of the faulted vein with an average movement determined by the two components. Repeated small dislocations practically result in a deflection of the vein (Fig. 33).

Overthrusts of great magnitude, such as are found in the Alps, may have had most important results as to the continuation in depth of ore deposits. As these dislocations may be measured in miles, it follows that whole groups of deposits contained in the overthrust portion of the strata may have been cut off entirely from their continuation in depth.[1] The relations of ore deposits to dynamic metamorphism is described in Chapter XXX.

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[1] B. Granigg, Ueber die Erzfuehrung der Ostalpen. Leoben, 1913.

CHAPTER X

OPENINGS IN ROCKS

Chemical processes and alteration in general may go on in a rock without cavities other than pore space and capillary or sub-capillary openings. Such processes are, however, metamorphic rather than metasomatic; they simply effect a mineralogical rearrangement without much chemical change; the composition of the rock remains constant. The formation of epigenetic mineral deposits usually implies a considerable addition of foreign material by solutions and these solutions must be guided to the place of deposition by open spaces, such as fissures, joints, or cracks. As a matter of fact the great majority of mineral deposits were formed where the path of the solution was prescribed by openings in the rocks other than those of ordinary pore space. After the solutions have gained access to the rock they may of course enter the pores and capillary openings and effect metasomatic changes. The sizes of capillary and subcapillary openings are given on p. 30.

The discussion which follows relates mainly to openings of supercapillary size. Such openings are chiefly found in the zone of fracture (p. 72). Few of our mineral deposits have been formed at depths much greater than 15,000 feet. Small openings may, however, exist in hard rocks at a distance below the surface much greater than the figure just indicated (p. 73). The possibility is, therefore, shown that solutions from great depths may gain access to the upper zone of fracture.

ORIGIN OF OPENINGS

Rock cavities may originate in various ways:

1. By the Original Mode of Formation of the Rocks.—Many volcanic flows contain abundant gas pores, or blow holes produced by the expansive force of gases escaping from the magma. Zeolites and calcite, sometimes with native copper, often accumulate in these pores, and such rocks are usually termed "amygdaloids" and the filled cavities "amygdules" (Fig. 34). Some sandstones and conglomerates contain much pore space in which solutions may deposit ores or other substances.

2. **By Solution.**—Solution cavities are found mainly in easily soluble rocks, such as limestone, dolomite, gypsum, and salt. Posepny justly maintains that the solvent power of water suffices to produce long galleries or passages in rock salt and mentions several examples.[1] Joints in limestone are often irregularly enlarged by solution and when subsequently filled with ores such cavities are known as *gash veins* or *pipe veins*. Caves in limestone are likewise made by atmospheric water of the upper circulation, containing dissolved carbon dioxide. Such caves are generally formed above the ground-water level in the zone of oxidation, though cases are known which suggest that the process

FIG. 34.—Photomicrograph of basalt showing blowholes filled with chlorite, calcite and native copper. Black areas represent copper. *After Volney Lewis.*

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[1] Genesis of ore deposits, 1902, p. 20. Pub. by Am. Inst. Min. Eng., New York.

can go on also below this level. Caves occur in all limestone regions and are sometimes of enormous extent; the Mammoth Cave of Kentucky has passages more than 40 miles in length and has been formed by the removal of millions of cubic yards of rock. The extent of caves is generally determined by faults and dislocations, and rock openings on a smaller scale are usually determined by the prevailing joint systems. The breaking in of caves near the surface produces the "sink-holes" so characteristic of certain limestone plateaus. Both caves and sink-holes have a certain importance in the origin of the class of zinc-lead deposits common to many limestone areas, and caves of dissolution in the oxidized part of ore deposits in limestone are sometimes the receptacles for a great variety of secondary minerals. The floors of caves are usually covered with red "cave earth," a residual deposit of silica, kaolin, limonite, etc., derived from the less soluble constituents of the limestone. Deposits of bat guano and nitrates are sometimes found in caves. Small solution cavities are often found in more resistant rocks that have been exposed to hot solutions of great solvent power.

3. **By Fractures of Various Modes of Origin.** (*a*) *Contraction Joints Produced by Tensile Stress in Igneous Rocks.*—When magmas congeal to igneous rocks tensile stresses which result in fissures and joints are developed. This is best exemplified in effusive rocks, which often show regular columnar structure and which are always full of irregular joints and cracks. No doubt these open spaces may guide metal-bearing solutions. In the literature many authors attribute fissure veins in effusive rocks to contraction, but usually without sufficient reason. The tensile stresses cannot produce long fissures with regular strike and dip.

According to the views of many geologists, smaller irregular veins in dikes or other intrusive rock masses fill contraction fissures. This explanation has been advanced for the horizontal tin-bearing joints in the Zinnwald granite, Saxony, and for other similar "stockworks;" also for the so-called "ladder veins," which are short transverse fissures in dikes, usually extending only from wall to wall. Well-known examples of this kind in Telemarken, Norway,[1] have been described by Vogt; in Victoria, Australia,[2] by Whitelaw; and at Beresowsk, in the

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[1] Zur Klassification der Erzvorkommen, *Zeitschr. prakt. Geol.*, 1895, p. 149.

[2] *Mem.*, Geol. Survey Victoria, vol. 3, 1905, p. 11.

Ural Mountains,[1] by Rose, Helmhacker, Karpinsky, Posepny, and Purington. In places, however, the transverse fissures may extend over the contact into the wall rock or correspond to the general joint systems of the vicinity, a fact which throws some doubt on the correctness of the explanation given (Fig. 35).

(*b*) *Contraction Joints by Shrinking of Limestone when Changed to Dolomite.*—Dolomite is not uncommonly formed near certain metal deposits and it is possible that this process when carried on by rapidly moving solutions and in comparatively free space may result in openings suitable as receptacles for ore minerals.

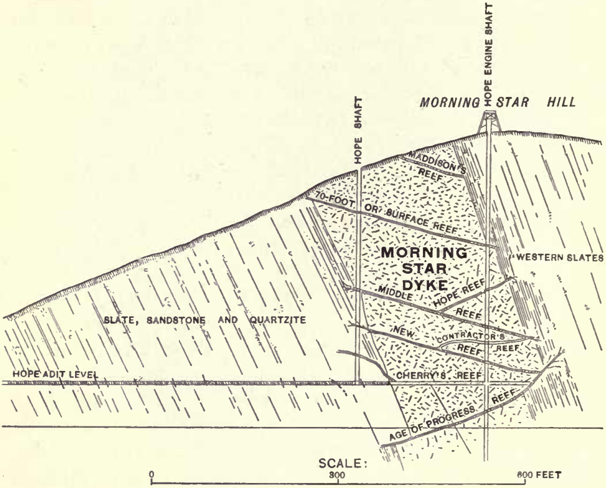

FIG. 35.—Section of Morning Star dyke, Woods Point, Victoria, showing ladder veins. *After O. A. L. Whitelaw.*

(*c*) *Expansion Joints Produced by Increase of Rock Volume.*—Peridotite upon change to serpentine near the surface and near fissures is believed to increase its volume greatly and such serpentine often breaks into smooth fragments. Extreme irregularity is a characteristic of all expansion joints and they are of little importance in ore deposition.

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[1] *Guide,* Seventh Int. Geol. Congress, 1897, p. 42.

F. Posepny, *Archiv für prakt. Geologie,* vol. 2, 1895, p. 499.

C. W. Purington, *Eng. and Min. Jour.,* June 13, 1903.

(*d*) *Fissures Produced by Torsional Stress.*—The celebrated experiment by Daubrée[1] carried out by twisting a thick glass plate has shown that torsional stress may result in several systems of long and radiating fissures. This experiment has frequently been cited by geologists to explain divergent vein systems, but G. F. Becker has pointed out that such fissures do not follow approximate planes, like fissure veins, but are decidedly curved and warped. Becker[2] regards torsional stress as a system of tensions.

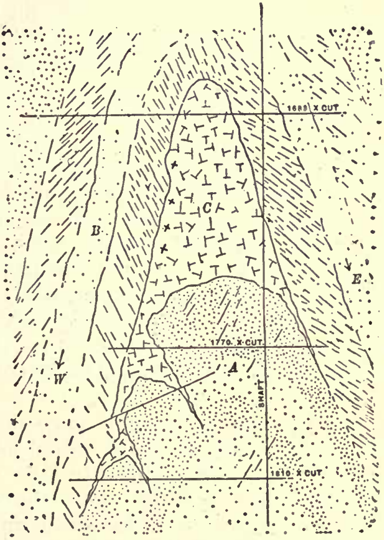

FIG. 36.—Section through a saddle reef, Bendigo, Victoria. *A*, Sandstone; *B*, shaly sandstone; *C*, gold-bearing quartz. *After T. A. Rickard.*

(*e*) *Openings Produced by Folding of Sedimentary Rocks.*—The bedding planes of sediments are primary structures which often serve as ducts for metal-bearing solutions. Better passageways for such solutions are provided when a series of sediments of unequal resistance is folded. A sandstone, for instance, will accommodate itself to bending with difficulty and will easily break at

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[1] Études synthétiques de géologie expérimentale, Paris, 1879, p. 316.

[2] The torsional theory of joints, *Trans.*, Am. Inst. Min. Eng., vol. 24, 1894, pp. 130–138.

anticlines or synclines, whereas softer shales will bend without breaking; the same process may cause a slipping between the various members. Such tensional stresses may then easily produce open cavities. The quartz-filled so-called "saddle reefs" of the gold mines of Bendigo and other places in Victoria are believed to have been formed in this maner by tensile stresses, but they are also accompanied by irregular masses or "makes" of quartz which fill spaces of discission across the beds (Figs. 13 and 36).

(*f*) *Openings Produced by Shearing Stress under the Influence of Gravity.*—In many disturbed regions the rocks are broken by normal faults along which the various blocks have settled down under the influence of gravity. Such normal faulting is especially characteristic of regions which do not bear evidence of strong compressive stress. Step-faulting is common and friction breccias and crushed zones frequently follow the faults; the open spaces, more or less continuous, offer good paths for the circulation of water if above the "level of discharge;" the fault planes are often long and regular. But in spite of all this, mineral deposits, except some of purely surface origin, are not common along such faults. At Clifton, Arizona, for instance, faults are abundant, but the copper deposits do not ordinarily appear in them. There are exceptions, however. At Creede, Colorado, a gold-silver vein occupies an important fault fissure, and similar cases are known from the silver-lead veins of the Harz Mountains in Germany.

In volcanic regions, such as Silverton and Cripple Creek (Fig. 37), in Colorado, systems of nearly vertical fissure veins contain rich deposits. They have obviously little connection with the main structural features of the country, the dislocations are usually small, and the veins were formed shortly after the close of volcanic activity. F. L. Ransome[1] believes that these fissure systems were generated by stresses resulting from slight vertical movements or settling, following an enormous transfer of volcanic material from an intratelluric to a superficial position. Vertical upthrusts of underlying magmas may have caused faulting accompanying or following vein formation.[2]

(*g*) *Openings Produced by Compressive Stress.*—In contrast to recently congealed lavas, the rocks which have formerly been far

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[1] F. L. Ransome, *Bull.* 182, U. S. Geol. Survey, 1901, p. 66.

[2] J. E. Spurr, Relation of ore deposition to faulting, *Econ. Geol.*, vol. 11, 1916, pp. 601–622.

below the surface of the earth but which have been exposed by erosion are usually traversed by more or less regular joint systems, persistent over large areas. While some of these joint systems may be caused by the inherent texture of the rock, they are in most cases the effect of compressive stress. Closely spaced joint systems form transitions into slaty cleavage, and recrystallization of minerals takes place by preference along these planes. In ex-

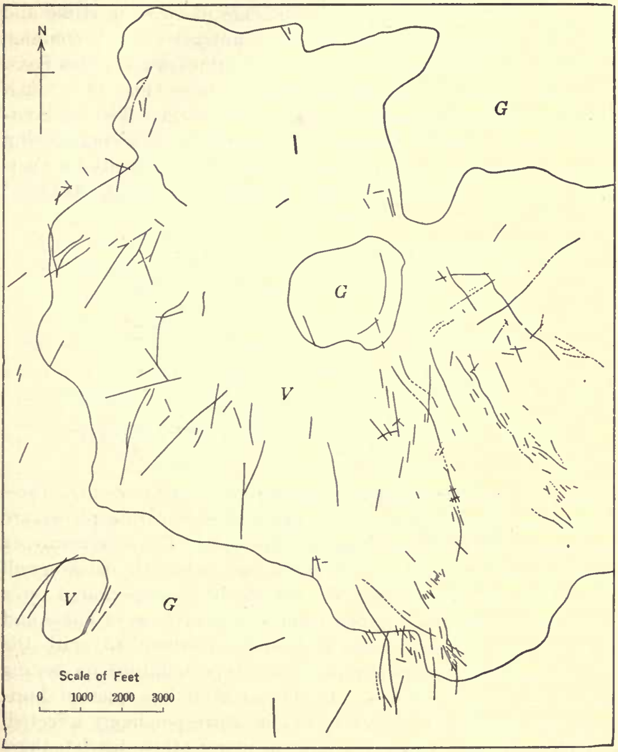

FIG. 37.—Plan of the principal veins of the Cripple Creek district, Colorado, showing a roughly radial distribution. *G*, granite and gneiss; *V*, tertiary volcanic rocks. *After Lindgren and Ransome, U. S. Geol. Survey.*

treme cases fissility or cleavage in very thin laminæ develops. Joints and cleavage present narrow paths for mineralizing solutions and ore deposits are often determined by their direction.

There are all transitions from joints to fissures along which perceptible movement has occurred. In many districts the fissures which have received the ores are identical in strike and dip with the joint systems of the country rock. A common condition is that two sets of veins and joints occur which have the same strike, but dip in opposite directions (Fig. 38). Such vein systems are termed conjugated fractures. The explanation of such joints, fissures, and occasionally accompanying schistose structure is furnished by certain experiments by Daubrée[1] and by the mathematical deductions of G. F. Becker.[2]

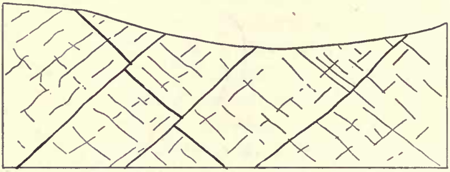

Fig. 38.—Vertical section of a conjugated system of fractures.

These show that compression develops two systems of fractures along the planes of maximum shear; these shearing planes are inclined to the direction of maximum stress. The accompanying dislocations will largely be reverse faults in which the hanging wall has relatively moved upward. In Daubrée's experiment on a mass of beeswax and resin two conjugated systems of joints and fissures were formed, making an angle of about 45° with the line of pressure; similar results have been obtained by testing cubes of building stones. If the stress is not exerted horizontally the dip of the veins will be correspondingly affected. At Grass Valley, California, and in many other districts there are two such conjugated systems of fissures which have been filled with ore.

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[1] Études synthétiques dc géologie expérimentale, Paris, 1879, p. 316.

[2] Finite homogeneous strain, etc., *Bull.*, Geol. Soc. America, vol. 4, 1893, p. 13.

C. K. Leith, Structural Geology, New York, 1913, p. 16.

In the locality just mentioned the majority of the dislocations are small, but tangential stresses sometimes produce great dislocations. The Mother Lode of California, a vein system nearly 100 miles in length, is believed to represent a reverse fault or system of faults with considerable throw.

When rocks are recrystallized in the deeper zones of the earth's crust they may become so plastic that deformation by rupture cannot take place. The growth of crystals then probably takes place predominantly in a plane perpendicular to the stress and a close schistose structure like that in many gneisses may develop which offers scarcely any interstitial space available for the circulation of solutions.

If the fissures were perfect planes it would be difficult to conceive of open spaces along them except by tensional stresses pulling the walls apart; but as they are not, movement along them tends to produce a series of openings, alternating with numerous touching points. As a matter of fact the mode of mineral deposition shows that open spaces existed and that they sometimes were large, in exceptional cases even 20 feet or more in width. In mine workings in hard rock old stopes frequently remain open for an indefinite length of time, and it is probable that such large open spaces may exist down to a depth of at least several thousand feet. Moreover, it is to be remembered that at the time of deposition the fissures were filled by water under a pressure at least equal to that of the hydrostatic column. The depositing solutions emanating from magmas under conditions of far stronger pressure may even have made way for themselves in the manner of an igneous dike or pegmatite vein, actually forcing the rocks apart. Some of the phenomena of deep-seated veins are difficult to explain on any other assumption.

Gaping fissures are not, however, necessary for the circulation of solutions. Water may ascend along a number of closely spaced fissures—usually called a sheeted zone—in which very little open space exists. But in this case mineral deposition is usually effected by replacement. The solutions are forced into the adjoining rock and transform its minerals into ore.

The stresses set up in a mass consisting of various rocks are extremely complex and it may only be possible to ascertain the dominant mode of fracturing.

Force of Crystallization.—Minerals crystallizing from solutions exert a certain pressure on the walls which confine them.[1] Many geologists have held that this force is sufficient to enlarge cavities along fractures and thus make room for mineral deposits. There is strong evidence in the structure and texture of veins which is unfavorable to such a view, except where conditions of light load prevail as near the surface or near open spaces. Some curious phenomena in regard to inclusions of rocks in veins may find their explanation by the action of this force, for in fissures filled with solutions a comparatively slight force might suffice to detach fragments from the walls.

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[1] G. F. Becker and A. L. Day, The linear force of growing crystals, *Proc.*, Washington Acad. Sci., vol. 7, 1905, pp. 282–288.

S. Taber, Pressure phenomena accompanying the growth of crystals, *Proc.* Nat. Acad. Sci., vol. 3, 1917, pp. 297–302.

S. Taber, *Am. Jour. Sci.*, 4th ser., vol. 41, 1916, p. 535.

CHAPTER XI

THE FORM AND STRUCTURE OF MINERAL DEPOSITS

The form of ore deposits is always important, for the mining methods used for a body of irregular outline must, for instance, be very different from those for a tabular vein. In the great majority of deposits the form is rudely tabular, for they usually follow the planes of dislocations or tabular dikes or the bedding of sedimentary rocks. Great weight was formerly attached to the form, both in empirical classification and in genetic interpretation. At present the tendency is to regard form as largely accidental, and to place more emphasis on the mineral association.

A convenient and fundamental though not strictly genetic classification divides mineral deposits into *syngenetic*, or those formed by processes similar to those which have formed the enclosing rock and in general simultaneously with it; and *epigenetic*, or those introduced into a pre-existing rock.

Syngenetic Deposits.—The syngenetic deposits include the magmatic segregations or accumulations of useful minerals formed by processes of differentiation in magmas, generally at a considerable depth below the surface. Their form may be wholly irregular or roughly spherical, but more often they are rudely tabular or lenticular, and they are usually connected by transitions with the surrounding rocks. They are either wholly enclosed in the igneous mass, or lie along its margins, or, in some cases, form dikes or offshoots from a deep-seated reservoir. The last class of ores may be called epigenetic with reference to the rocks incasing the dikes. The width and thickness of these deposits may range from a few inches to several hundred feet, and in rare cases, their length may exceed one mile. Masses of chromite in peridotite or titanic iron ore in anorthosite, furnish examples of this type.

The syngenetic deposits also include sedimentary beds; they have, as a rule, a tabular or sheet-like form; they are horizontal if not disturbed, but are frequently folded and faulted. Parallel to their bedding their extent may be counted by miles, as in the case of the Clinton hematite ores of the Appalachian region, or the French and German limonite beds; nevertheless, each bed usually thins out in wedge-shaped form and may be replaced by others at a slightly different horizon. In deposits of metallic ores the thickness is rarely more than 20 feet and this may include intercalated beds of barren material. Coal beds, especially those of lignite, or brown coal, may attain a thickness of 100 feet or more. Beds of rock salt, anhydrite, and gypsum are in some cases several hundred feet thick. In all sedimentary deposits displacements and folding may locally produce an appearance of great thickness. In plastic material like rock salt such deformation is especially effective.

Epigenetic Deposits.—The epigenetic deposits have various forms, but among those which follow fissures the tabular or sheet-like form is most common. Deposits concentrated in the zone of weathering are often extremely irregular and of limited extent, and several of them are usually found in close proximity. Some hematite ores, like those of the Mayari district in Cuba, which are developed by the weathering of serpentine, may form superficial sheets of great extent.

Replacement deposits in limestone are extremely irregular, although their form as a whole is often dependent upon the bedding, the fissuring, or the contact with other rocks. They are seldom large, but in a few cases, like the galena deposits in southeastern Missouri or the zinc blende deposits in the Joplin region of the same State, they may be followed at a general horizon for several miles.

The ore deposits in metamorphic rocks which have undergone strong mechanical deformation and chemical changes usually assume lenticular form, and the occurrence of successively overlapping lenses is particularly characteristic. In these deposits a steep dip is a common feature, but the main trend of the ore-body in the plane of its strike is usually not in the direction of the dip.

The *strike* of a tabular or lenticular deposit is the direction of a horizontal line in the plane of the deposit, measured with reference to a meridian.

The *dip* is measured by the vertical angle between a horizontal plane and the plane of the deposit. Complementary to the dip is the *hade* or *underlie,* which is measured by the angle between the vertical and the plane of the deposit.

The *plunge*[1] (Fig. 39) of an ore-body is the vertical angle between a horizontal plane and the line of maximum elongation of the body. In lenticular ore-bodies in metamorphic rocks which have undergone strong mechanical deformation, the plunge is an important factor, and often it is determined by the direction of the cleavage or schistosity. In fissure veins the *pitch* of the ore shoot is usually defined as the angle between its axis and the strike of the vein, and it is measured on the plane of the vein[2] (p. 184).

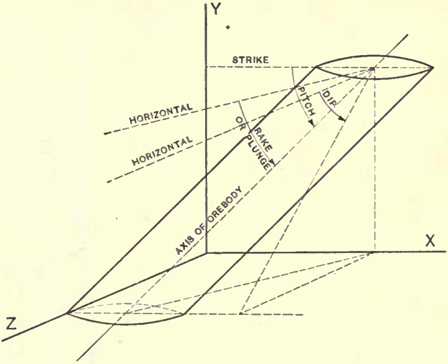

Fig. 39.—Stereogram illustrating strike, dip, pitch and plunge of an ore-body.

Spacial Relations of Veins.—Veins are tabular or sheet-like masses of minerals occupying or following a fracture or a set of fractures in the enclosing rock; they have been formed later than the country rock and the fractures, either by filling of the open spaces or by partial or complete replacement of the adjoining rock, or most commonly by both of these processes combined.

Such alteration or replacement does not ordinarily extend far from the fissure. In regions where the vein-forming solu-

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[1] Called "pitch" or "rake" by many authors.

[2] See discussion in *Trans.,* Am. Inst. Min. Eng., vol. 39, 1908, pp. 898–916.

tions have acted with unusual intensity a partial alteration may extend from the deposit over considerable areas.

No sharp distinction can be drawn between the filled veins and replacement veins. If open spaces are available the metalliferous solutions which formed the veins in most cases found it easier to deposit their load in these spaces than to replace the country rock. Quartz is more likely to be deposited in the open paths, and likewise most of the heavy metals, unless the country rock is one particularly adapted for replacement, such as limestone. Gases like carbon dioxide and hydrogen sulphide penetrate the wall rocks with ease.

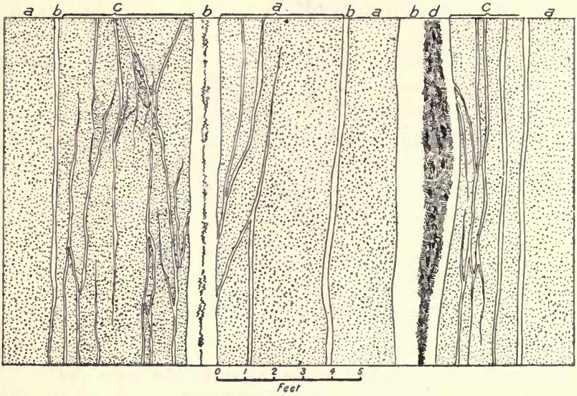

Fig. 40.—Section of Silver Crown lode, Silverton, Colorado, showing lode structure. *a*, Andesite; *b*, quartz; *c*, andesite and quartz stringers; *d*, ore. *After F. L. Ransome, U. S. Geol. Survey.*

Many veins correspond closely to the old definition of a "true fissure vein," in which the ore occupies the once open spaces along the fracture, with some alteration spreading into the wall rocks. Of such character are the majority of the gold-quartz veins of California and many other occurrences. When the fissures are very small they are referred to as veinlets or seams, and all transitions to a slight mineralization of joint planes are found. The walls may be smooth and separated from the vein material by a *clay gouge* or the filling may closely adhere to the country rock. In the latter case the vein is said to be *frozen to the walls.*

Instead of a single break we may have a fracture consisting of a number of approximately parallel fissures, irregularly connected and spaced over a considerable width, which may attain 100 feet or even several hundred feet. These large fracture zones, when filled with ore and partially replaced country rock, are called *composite veins* or *lodes* (Fig. 40). The Comstock lode in Nevada illustrates this occurrence; its width in places amounts to several hundred feet.

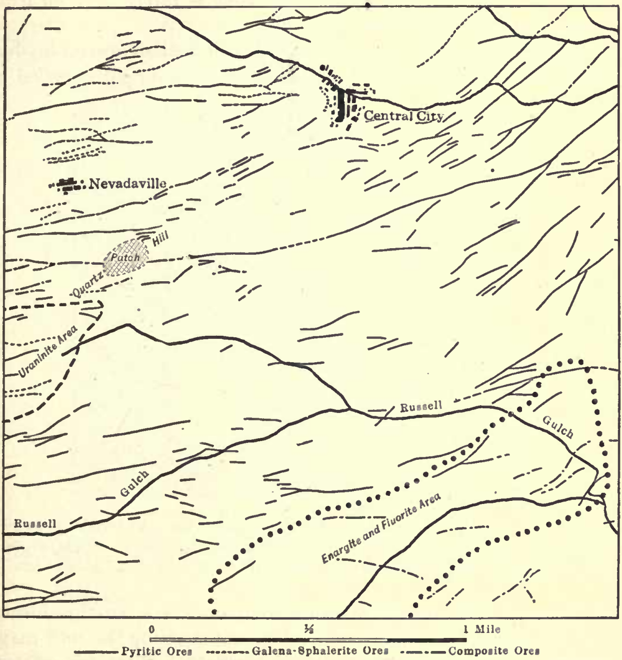

Fig. 41.—Map showing veins of Central City, Colorado and vicinity. *After E. S. Bastin, U. S. Geol. Survey.*

Lodes often contain two systems of fractures, intersecting at an acute angle, as shown roughly on Fig. 40. This is sometimes referred to as *hammock structure.* A number of adjacent parallel veins are called a vein system. If connected by diagonal veins the term *linked veins* (Fig. 41) is used.

When the fractures are closely spaced and parallel we speak of a *sheeted zone* or a *shear zone* (Fig. 42). Many of the Cripple Creek veins form good illustrations of this mode of occurrence. The width of a sheeted and mineralized zone is rarely over 50 feet and ordinarily much less.

A mass of rock irregularly fractured in various directions by short fissures along which mineralization has spread is called a *stockwork.* Gold-quartz deposits sometimes assume this form; each seam in the several joint systems intersecting the rock may contain a thin but often strongly auriferous sheet of quartz; the mass may be mined as a whole, furnishing low-grade ore. In deeply weathered regions the upper parts of such deposits may be sufficiently disintegrated to be washed by the hydraulic method. In California such mines are called "seam diggings."

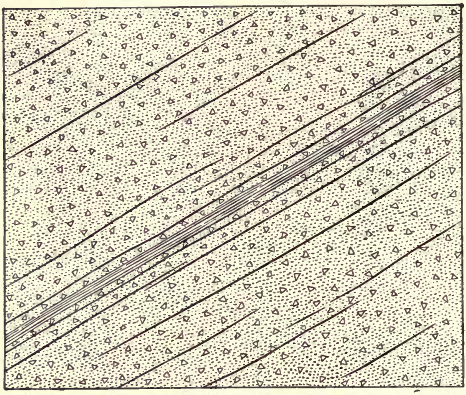

FIG. 42.—Section of the Howard vein, Cripple Creek, Colorado, showing a sheeted zone. Ore follows the close sheeting in the center. Scale, 1 inch equals 13 feet. *After Lindgren and Ransome, U. S. Geol. Survey.*

A shattered zone cemented by a network of small non-persistent veins is called a *stringer lead* or *stringer lode.*

Sometimes ore deposits are wholly irregular brecciated masses, the ores filling the interstices between the fragments. Again, the breccia may be localized at the intersection of two fractures and a *pipe-like* deposit will be formed, the ore cementing the fragments. Or again, ore deposition may have proceeded in a volcanic vent filled with fragments of rocks due to explosive action. Of such character was the celebrated Bassick deposit in Custer County, Colorado.

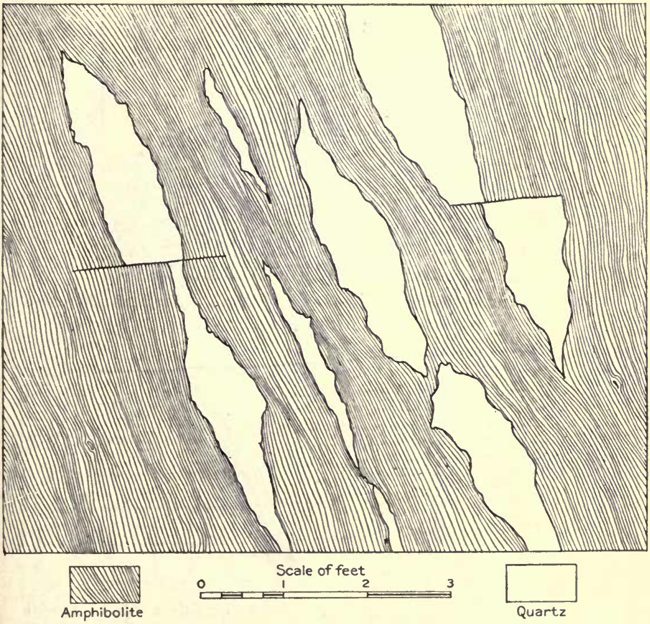

Fig. 43.—Vertical section of Schlegelmilch quartz vein, South Carolina, showing lenticular vein structure in schist with offsets along joint-planes. *After L. C. Graton, U. S. Geol. Survey.*

Brecciation, shattering, and mineralization often follow lines of weakness along dikes; in such cases, illustrated by the Douglas Island mines in southern Alaska, where a dike of diorite intrudes metamorphic clay slates, the mineralized dike is often referred to as a lode.

Ladder veins are deposits filling short transverse fissures sometimes occurring in dikes of intrusive rocks (see Fig. 35).

Lenticular veins (Fig. 43) are confined mainly to metamorphic schists and their form is sometimes caused by deformation of an older deposit; or again the lenticular shape may be due to stresses causing bulging of the schistose layers. It is common to find a number of short lenses of gold-bearing quartz, for instance, scattered along a certain line or zone. Their ends sometimes overlap.

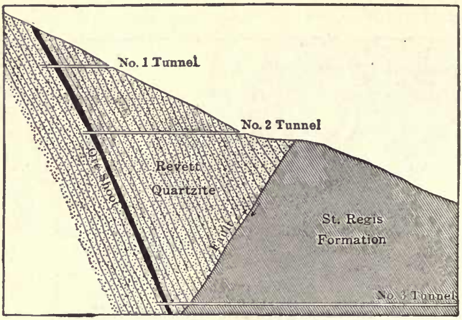

Fig. 44.—Section of Snowstorm bed-vein, Idaho. *After F. L. Ransome, U. S. Geol. Survey.*

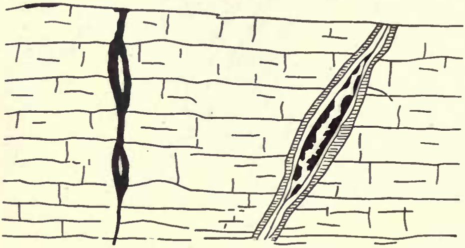

Fig. 45.—Vertical section of gash veins filled with galena (black), or with pyrite, zinc blende, and galena in order of deposition. Drusy cavities in center. Lead mines of Wisconsin. *After T. C. Chamberlin.*

Bed veins follow the bedding planes in sedimentary rocks (Fig. 44).

Gash veins are deposits filling non-persistent openings that are of fair width but soon cease when followed along strike or dip;

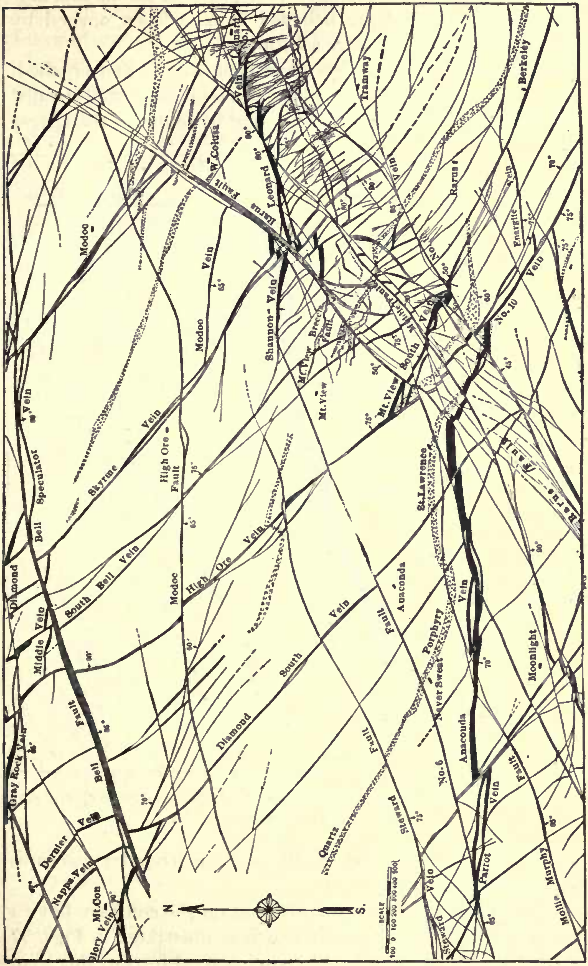

FIG. 46.—Horizontal section of part of Butte district, Montana, on level about 1,500 feet below surface. *After Reno Sales.*

they are particularly characteristic of deposits of galena and zinc blende in limestone and are believed to have been opened by tensional stress, often aided by solution.

Where soft sedimentary beds have been folded and crushed, irregular open spaces are more likely to result than well-defined straight fissures. In such rocks ores may be found in the spaces opened along anticlines and synclines or in irregular fractures breaking across such folds.

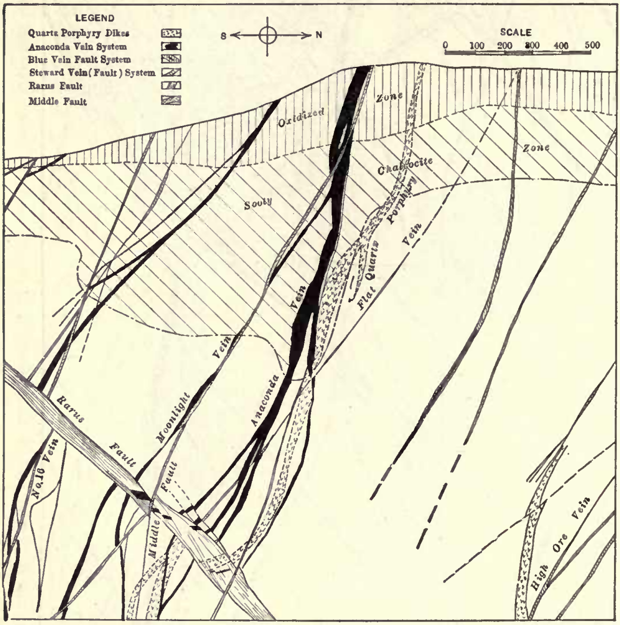

FIG. 47.—N.-S. section across Butte district, showing structure and ore zones. *After Reno Sales.*

Veins and lodes rarely occur single but on the contrary have a tendency to cluster in *vein systems* such as illustrated in Figs. 37 and 41. In some places may be found several intersecting vein systems of great complexity and differing ages and differing mineralization as, for instance, is the case in the great copper district of Butte, in Montana, illustrated in plan and section in Figs. 46 and 47.[1] The peculiar divergent fractures at the Leonard mine form what is sometimes called a horsetail structure. The veins at Butte are moreover in many places disrupted by later faults.

Veins in Relation to the Country Rock.—Veins crossing the bedding in stratified rocks are referred to as cross veins; those parallel to the stratification or schistosity are often called bedded veins or bed veins. Differences in the texture and hardness of the rocks traversed influence the form of the vein markedly. In hard dikes crossed by the vein the deposit often splits up into stringers, resuming its typical form beyond this barrier. In fractures formed under light load near the surface there is a great tendency to irregularity and brecciation, especially in the hanging wall. Following G. F. Becker's proposal such may be called *chambered veins.* In a vein of strong dip there will also be a tendency for the hanging wall to settle accompanied by the development of minor vertical fissures. Such conditions were found, for instance, in the Comstock lode, Nevada (Fig. 165) and in the El Oro mines of Mexico (Fig. 159); at both places the vertical hanging wall veins were exceptionally rich, the richness being possibly caused by the impeded circulation of the depositing waters. Large masses of country rock included in the vein material are called *horses.* Frequently the vein follows a fissure along the walls of a dike; the lamprophyric dikes which are the last phases of batholithic intrusions are especially favored places for ore deposition.

Clayey and soft rocks are most resistant to the development of regular fissures; a fracture in hard rock will suddenly die out when encountering such material; many veins pinch immediately upon entering clay shales or masses of clayey gouge. One of the best examples of this is furnished by the veins of Rico, Colorado, which do not extend through the whole sedimentary series in that district, but suddenly cease at a certain stratum of yielding, plastic rocks, termed the blanket, under which almost all the ore-bodies occur. A consequence of this peculiarity of fissuring is that in some regions rich ores are often found just below certain horizons of shale. In southern New Mexico a persistent Devonian shale plays this part of "indicator" (Fig. 70).

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[1] Reno Sales, Ore deposits of Butte, Montana, *Trans.,* Am. Inst. Min. Eng., vol. 46, 1913, pp. 1–109.

The vein solutions were arrested at this horizon and there deposited their load.

When a vein follows the contact between two formations, say between granite and andesite, we speak of it as a *contact vein.* The contact is usually caused by faulting movements in the plane of the fissure, and such veins are in no wise different from ordinary fissure fillings. They should not be confused with contact-metamorphic deposits, which belong to a separate class.

Vein Walls.—In a simple filled fissure vein we have well-defined foot and hanging walls, which often are smooth surfaces and represent a single fissure opened by a small or large movement along its slightly curved plane. In a replacement vein the fissures are comparatively tight and in most cases appear to have been formed under stronger compressive stress that reduced the open spaces to a minimum. The vein-forming solutions were forced into the country rock, and the ores formed by replacement gradually merge into unaltered rock. In such cases we may find a single fissure plane with ore on both sides and not limited by any well-defined walls. The exact limits of commercial ore can be found only by assay and are often spoken of as "assay walls."

In a composite vein or lode or in a sheeted zone there may be several smooth walls and if no cross-cutting is undertaken there is danger that parallel ore-bodies separated by sheets of country rock may be overlooked.

Outcrops.—The character of the outcrop of a vein, or in fact of any deposit, is determined by the predominant minerals and by the prevailing surface conditions. In regions of long-continued rock decomposition and inactive erosion, as, for instance in some of the Southern Appalachian States, even the most resistant outcrops may be reduced by weathering and nothing but fragments scattered over a wide area may be visible at the surface.

Under conditions of fairly active erosion veins with predominant quartz stand out prominently and can be easily traced.

On the other hand, veins with carbonate gangue are likely to weather more rapidly than the surrounding rock, and the deposits may be indicated by little depressions or by notches in the ridges. Where the sulphides are abundant, their oxidation is conspicuously reflected in the outcrops. Deposits of mingled quartz and sulphides then form prominent outcrops of limonite and residual quartz; this is the *gossan* of the Cornish, the *ironstone* of the Australian, the *eiserner Hut* of the German, and the *colorados* of the Spanish terminology. More details in regard to the weathering of ore-deposits are given in Chapter XVIII.

Length and Depth of Veins.—Where veins follow great dislocations their length may be considerable. One of the more recent veins of Freiberg, Saxony, called the Halsbrücker Spat, has been followed for almost 5 miles. Some of the lead-bearing veins in the Harz Mountains, Germany, are traceable for 12 miles. Exceptionally long single ore-bearing fissures are found in the Silverton quadrangle, San Juan region, Colorado; some of them are 5 miles long. Some of the Mother Lode veins in California can be traced for many miles. The longest single quartz vein known appears to be that of the Pfal, in the Bavarian Forest, which is said to be traceable in a straight line practically without interruption for 140 kilometers through the pre-Cambrian rocks.[1] The quartz is said to be barren of metals.

The great majority of single ore-bearing veins are short and their outcrops can rarely be traced for more than one mile; they do not, as a rule, occupy great dislocations, but rather subordinate fissures. The great dislocations are formed during mountain building by tangential stresses, whereas the ore-bearing veins are, as a rule, formed after epochs of igneous activity. In the Coeur d'Alene district, Idaho, for instance, the rich galena veins show little connection with the principal structural faults of the region and were probably not formed at the same time.

Veins do not necessarily continue to great depths. There are all kinds of fissures, some disappearing within a short distance below the surface, others continuing down to the greatest depths attained, or about 6,000 feet (Morro Velho, Brazil). Deep tunnels have been run to intersect veins of favorable appearance on the surface and have failed to disclose their continuation in depth. There is no definite relationship between depth and length of a fissure, though it is true that fissures showing strong movement and shattering are likely to continue to great depths. The ore-body may be limited in depth, while the barren fissure continues below it as strong as ever.

Bends and curves in strike and dip are common in veins, but

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[1] E. Suess, Das Antlitz der Erde, Leipzig, 1883, vol. 1, pp. 270–272.
W. von Gümbel, Geologie von Bayern, Cassel, 1894, vol. 2, pp. 461–464.

as a rule a vein retains its general angle of dip with remarkable persistence. The dip may be at any angle, but veins dipping from 50° to 80° are most common. The North Star vein at Grass Valley, California, is one of the best instances of a low-dipping vein of great length; with a dip of 20° it has been followed for 5,000 feet. Still flatter veins are called blanket veins and seldom are very persistent or uniform.

CHAPTER XII

THE TEXTURE OF MINERAL DEPOSITS

FILLING AND REPLACEMENT

Introduction.—The ore minerals and gangue which make up an ore deposit present various types of texture. The texture of an ore is dependent upon many factors. Space available for deposition, concentration and composition of the generating solutions, time, temperature, and pressure—all are of importance in determining the primary texture. Many changes take place in a deposit once formed; the secondary textures, so far as they are caused by solution and redeposition, are influenced by the same factors, and, in addition, deformation by pressure plays a most important rôle.

Texture of Deposits of Igneous Origin.—The ores consolidated from magmas have in general the texture of igneous holocrystalline rock. The principal minerals comprise chalcopyrite, pyrite, pyrrhotite, magnetite, chromite, and ilmenite. The texture is ordinarily coarse granular, hypidiomorphic; the chalcopyrite and pyrrhotite are rarely crystallized, but may contain phenocrysts of pyrite and magnetite, both of which are frequently developed with crystalline outlines. The ores may, of course, contain phenocrysts and anhedrons of other rock-forming minerals, particularly soda-lime feldspars, olivine, and pyroxene. Eutectic texture results if the magma was a eutectic mixture from which two minerals crystallized simultaneously after the manner of graphic granite. Approximation at least to such texture is shown by some intergrowths of magnetite and apatite.

If the ores have been subjected to dynamic metamorphism, granulation and metasomatic development of hornblende, garnet, biotite, and epidote in coarse or fine aggregates follow and the ore may acquire schistose structure.

Texture of Pegmatite Dikes.—The pegmatite dikes are believed to have been deposited by magmatic solutions of great fluidity and low temperature (about 600° C.). In many cases the pegmatites form transitions between igneous rocks and veins deposited by hot solutions. Their texture is coarsely crystalline, often drusy, and the minerals have a strong tendency to idiomorphic development. Large crystals are the rule, and sometimes they attain enormous dimensions; crystals of spodumene at the Etta mine, South Dakota, are 30 feet or more in length. Quartz crystals several feet long have been observed in these deposits. A rough tendency to crustification is often present, and the walls of the dikes are then lined with crystals of feldspar or mica.

Texture of Sedimentary Deposits.—Ores and minerals of sedimentary deposits are usually fine grained, and in many cases they have been deposited as colloids in which subsequent fine-grained crystallization has developed. Coarsely crystalline, allotriomorphic structure may develop in deposits consisting of calcite, salt, or gypsum.

In many cases the structure is clastic with development of new-formed minerals between the grains. Newly formed quartz, if present, nearly always assumes a microcrystalline or cryptocrystalline texture. Subsequent metamorphism is likely to enlarge the crystalline grains and result in coarser-grained ores.

Concretions.[1]—Concretions are rounded bodies of some mineral aggregate which are often found in shale and sandstone. Calcite, silica, siderite, pyrolusite, barite, pyrite, marcasite and limonite are among the minerals which most commonly form concretions. The structure is often concentric or radial. In some cases the stratification planes pass through the concretions, while in other cases they may bend around them. These structures are of some economic importance as regards ores of iron and manganese, especially, siderite, limonite and pyrolusite. They often have a center of a clastic grain or a fragment of a fossil shell or leaf. The concretions result from processes of solution and precipitation in soft or semi-consolidated sediments. Accidental precipitation, say of pyrite around decomposing organic material, may start the action and the laws of mass action and preferred growth of larger crystals continue the process. Concretions generally derive their substance from the surrounding rock. Sometimes the minerals simply fill pores and interstices; but in many cases the original substance may have been removed by metasomatic processes. Concretions are frequently

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[1] J. E. Todd, Concretions and their geological effect, *Bull.*, Geol. Soc. Am., vol. 14, 1904, pp. 353–368.

James Geikie, Structural and field geology, 1905, Chapter VIII.

altered with volume changes and development of cracks and interior cavities. When small, uniform and abundant they are called oölites. The oölities usually result from separation in colloidal solutions, often coupled with adsorption of electrolytes.[1]

The oölitic texture is characteristic of many deposits of calcite, siderite, calcium phosphate, limonite and psilomelane; pyrite rarely assumes this form. The oölites are often affected by later alteration and recrystallization.

Texture of Residual and Oxidized Deposits.—In the residual deposits of the zone of oxidation, the ore-bodies are usually very irregular in structure and texture. In large part they were deposited as colloids, which subsequently in part have developed fine-grained crystalline texture.

Earthy, clayey concretionary, mammillary, stalactitic, or pisolitic textures are common, the last being defined as a coarser development of the oölitic form. Coarser crystalline form is assumed by some minerals like calcite, barite, zinc carbonate, zinc silicate, and lead carbonate. Crustification or drusy structure is common in places. Quartz, where developed, is usually fine-grained or cryptocrystalline.

THE TEXTURE OF EPIGENETIC DEPOSITS

Primary Texture of Filled Deposits.—The epigenetic deposits are of manifold form and origin, but the majority of them result from aqueous solutions either by filling of open cavities or by replacement of surrounding rocks. Precipitation from complex solutions in open spaces takes place in a certain orderly succession, and the deposits therefore readily assume a banded texture; crystallization is facilitated by the open spaces, but the older crystals interrupt the development of the products of later crystallization. Hence a hypidiomorphic to panidiomorphic texture is most common.

Banding by deposition is called *crustification*, a term introduced by Posepny. In many classes of veins, whether banded or not, a drusy texture is common.

In deep-seated veins formed at a temperature but slightly lower than that of the pegmatites the texture is usually coarsely crystalline and massive; sometimes even drusy cavities are

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[1] For recent literature regarding the origin of oölites see Fortschritte der Min. Krist., u. Petr. Jena. 1913, p. 43.

lacking. Delicate and repeated banding is absent, but a coarsely banded or comb structure recalling that of the pegmatite veins is sometimes encountered. It is usually expressed by quartz crystals developing from the sides or by metallic minerals like tourmaline, wolframite, or cassiterite attached to the walls of the fissure.

In veins formed at intermediate temperatures a coarsely crystalline massive texture is most common; combs and rough banding by deposition are by no means unknown, especially where the deposit contains calcite or barite. In quartz veins the filling appears to have taken place rapidly and completely, so that the resulting ore consists of an irregular massive mixture of quartz and sulphides. That here too the deposition began from the walls is indicated by some occurrences of partially filled veins which form a loose aggregate of prisms. Any

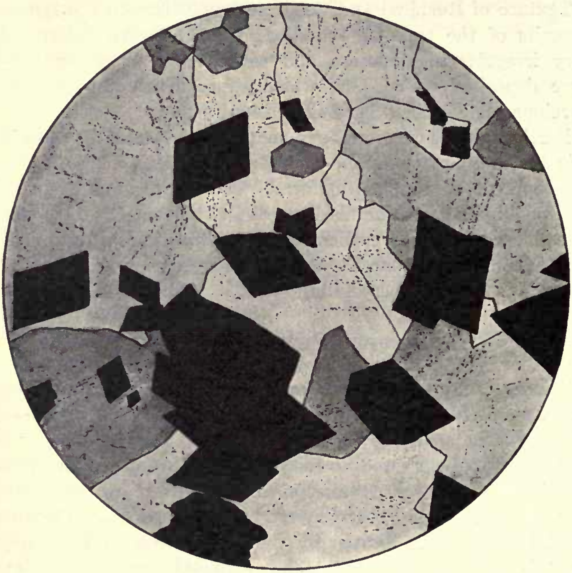

Fig. 48.—Thin section showing normal texture of quartz filling. Black, arsenopyrite; remainder, quartz with fluid inclusions. Magnified 52 diameters. Gold quartz vein, Grass Valley, California.

thin section of such quartz will usually show long crystals of earlier growth around which the later quartz has been deposited in large individuals (Fig. 48). Lines of inclusions often penetrate from one grain into another. These inclusions consist of aqueous solutions, often with small cubes or grains of transparent salts suspended in the liquid. Inclusions of carbon dioxide have been reported, but are extremely scarce. The optical continuity of the crystals or grains is often disturbed by a peculiar divergent "flamboyant" structure which appears to be of primary origin, and not caused by internal strains.

The sulphides are coarsely crystalline and sometimes roughly banded, parallel to the walls. Inclusions of country rock may be surrounded by concentric rings of sulphides, and a primary brecciated vein structure may result. Pyrite and arsenopyrite, both among the earliest minerals, have a strong tendency to crystal development, while galena, zinc blende, chalcopyrite, and tetrahedrite are much less commonly found with crystal faces.

A banded structure sometimes results from the filling of several closely spaced fissures. In quartz veins in fissile rocks a peculiar *book structure* may result from numerous parallel sheets of slate, alternating with quartz. It has been thought that this and other features difficult to explain by the assumption of open cavities are due to the opening of spaces by the *force of crystallization*. Such views have been expressed by E. Suess, W. O. Crosby, E. J. Dunn, S. Taber, and others. It is improbable that crystallization could have opened the cavities. More likely they were supported by the strong pressure of magmatic waters. But within such spaces a slight force exerted by crystallization could readily detach fragments of shale from the walls.

Stalactites are unknown in deposits formed at high or intermediate temperature.

In veins formed at lower temperatures and comparatively shallow depths crustified and drusy forms and fine granular texture predominate. The quartz filling is usually fine-grained, ranging to cryptocrystalline and microcrystalline near the surface.

The sulphides are found in small crystals or small anhedrons; large crystals of pyrite, so common elsewhere, are rarely found in these veins. On the other hand, where calcite, dolomitic carbonates, rhodochrosite, fluorite, or barite are gangue minerals the crystals may be much larger than those found in other deposits. An example is furnished by the magnificent crystals of calcite at Joplin, Missouri, and here galena also appears in unusually large, well-developed individuals.

Symmetrical and delicate crustification is often associated with large drusy cavities. Brecciated structure of primary origin is common.

Secondary Textures and Structures of Filled Deposits.—Crushing and brecciation of the early minerals are extremely common; indeed, few veins are entirely free from it. Repeated opening of fissures (Fig. 51) and the deposition of new generations of vein material often take place and the cementing ore may be enriched at the expense of the older generations.

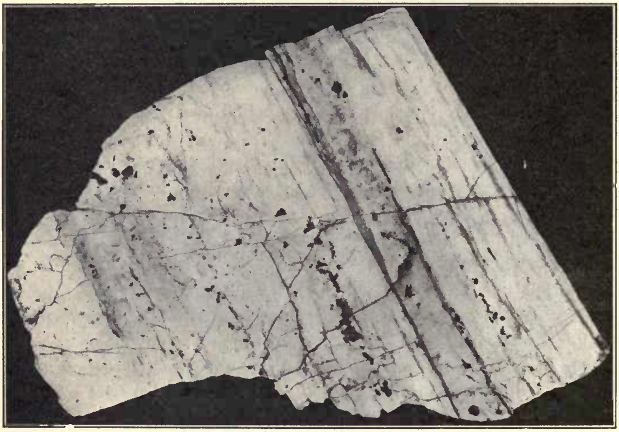

Fig. 49.—Specimen of quartz from Nevada City, California, showing ribbon structure by sheeting. Two-thirds natural size.

A banded or sheeted structure often results from the development of shear planes in the old filling; examples of this are seen in many gold-quartz veins of California (Figs. 49 and 50). Along these shear planes the quartz is deformed and granulated, and gold may be deposited along them by processes which may be called secondary, though, as a rule, they take place shortly after the vein formation. The shearing stress exerted either before or after the filling may affect the walls of the vein and render them close-jointed or even distinctly schistose.

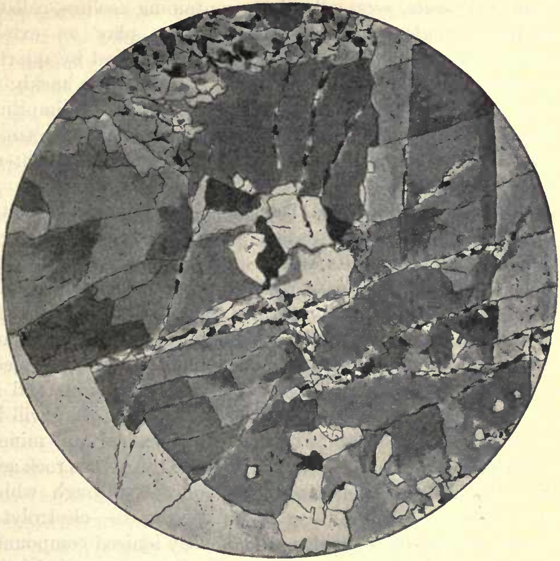

FIG. 50.—Thin section of vein quartz from Nevada City, California, showing crushing and incipient ribbon structure. Magnified 15 diameters. Crossed nicols.

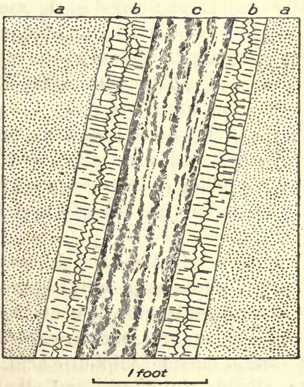

FIG. 51.—Cross section of Japan vein, Silverton, Colorado, showing structure produced by repeated opening of original fissure. *a*, Country rock; *b*, quartz; *c*, ore. *After F. L. Ransome, U. S. Geol. Survey.*

In some deposits, especially those containing zeolites, calcite, or barite, secondary replacement processes play an extensive part. A vein filled by calcite may be replaced by quartz, which then plainly shows its secondary nature by its hackly or lamellar texture, casts of cleavage, pieces of calcite, or imprints of cleavage lines. Such pseudomorphic textures are sometimes accompanied by a marked enrichment of the metallic content of the deposit.

Metasomatism in Mineral Deposits.—The nature of metasomatism or replacement has already been described on pages 26 and 69. Many deposits have been formed by solutions containing various salts and gases and capable of attacking certain kinds of rocks. Guided by fissures or other open ducts the solutions deposit part of their load in the open supercapillary spaces whenever supersaturation takes place; thus is produced the filling of fissures. As almost all rocks are porous and as the solutions are frequently under heavy pressure they will be forced into the rocks and will produce chemical and mineralogical changes in them. At the same time the porous rock acts undoubtedly as a semi-permeable membrane through which various substances will diffuse at differing rates—electrolytes and gases most easily, colloids and difficultly ionized compounds very slowly. Thus any vein will usually be accompanied by a strip of altered country rock in which the solutions have effected certain metasomatic changes. The minerals in the open fissures will ordinarily differ from those formed in the metasomatic zone. We may find, for instance, a quartz filling with various sulphides and gold, while the minerals developed in the country rock consist of pyrite, sericite and calcite with little if any gold. In some cases no perceptible alteration may be observed in the country rock. The only difference between a filled vein accompanied by metasomatism and a so-called replacement deposit is that in the latter the filling of the narrow open spaces is negligible and the bulk of the ore has been formed by metasomatic processes.

Metasomatic Processes.—In a solid rock replacement may be caused by many kinds of solutions the only requirement being that some or all of the rock minerals must be unstable in the penetrating fluids. The usual substances, most active in aqueous solutions, are oxygen, carbon dioxide, sulphuric acid, ferric sulphate, hydrogen sulphide, alkaline sulphides, and alkaline carbonates.

Replacement may occur at all temperatures above the freezing point of the solution and below the melting point of the rock; it is naturally most effective in hot solutions. Replacement may proceed at any pressure. It may be effected by the ordinary surface waters, by sea water, by hot ascending waters and by magmatic emanations whether gaseous, fluid or above the critical temperature.

There is no rock that is proof against replacing natural solutions of some kind. Limestone and dolomite are most easily replaced and even at ordinary temperatures, for instance, by iron carbonate (siderite) or by zinc carbonate (smithsonite). Granite, diorite, and in fact all igneous rocks are also subject to replacement. Even quartzite, slate and aluminous shale may be replaced by other minerals though they are more resistant than others. Replacement by sulphides may to some extent take place at ordinary temperatures (for instance chalcocite replacing pyrite) but large deposits of sulphide ore are usually formed by hot solutions.

Mode of Replacement.—As pointed out in previous chapters replacement is effected by concentrated solutions filling capillary openings of extremely small size (sheet openings larger than 0.0001 mm., p. 30), which are just above or below the limit of microscopic visibility. Cases have been noted when replacement begins from a crack doubtless filled with a film of solution and connecting a series of just visible fluid inclusions. Solution and precipitation go on practically simultaneously dependent upon the constantly changing equilibrium, the supply of solvent and the facility of escape for the dissolved material. Two or several minerals may be dissolved at the same time to make room for the new as in the replacement of shale by a pyrite crystal. The volume of the rock remains constant, held by pressure. The moment a place is available some mineral will separate out from the concentrated solution. This law fails to apply in free crystals or when rock pressure can be overcome by the force of crystallization, or when a solid is replaced by a gel, or when the solutions circulate so rapidly that there is a strong balance in favor of solution. As crystal grains develop they will exert a different amount of pressure in various directions thus facilitating solution in the direction of greatest pressure. The development of crystals in the host mineral is a result of this action.

The power of crystallization of the different minerals varies greatly, for some are found only as anhedrons in metasomatic rocks, while others always assume their crystal form. The following list gives the relative power of crystallization in solid rocks of some minerals, as beginning with those of strongly emphasized individuality: Rutile, tourmaline, staurolite, arsenopyrite, pyrite, magnetite, barite, fluorite, epidote, pyroxene, amphibole, siderite, dolomite, albite, mica, galena, zinc blende, calcite, quartz, orthoclase.

When a crystal has ceased to grow solution may still continue parallel to its faces. As no more material for the crystal is at hand the voids are immediately filled by the next precipitate available. Thus are explained the thin films of quartz or calcite which so frequently surround metasomatic pyrite crystals.

When taking place under the law of constant volume replacement cannot ordinarily be expressed by the simple chemical formulas[1] usually given. The reactions are likely to be more complicated.[2]

In metasomatic processes gangue minerals like sericite, calcite, siderite, barite and fluorite replace all silicates. Ferromagnesian silicates will be attacked first, then the soda-lime feldspars, lastly orthoclase and albite. The degree of attack on quartz depends probably on the amount of alkaline carbonates in the solution. All sulphides replace all silicates as well as quartz. (Fig. 52). Sulphides and sulphosalts readily replace other sulphides. A succession common in many ores is (1) pyrite (oldest), (2) chalcopyrite, (3) galena and zinc blende, (4) sulphosalts (like tetrahedrite). Any of the later minerals may replace any of the earlier products (Fig. 53). Our knowledge of these manifold replacements have been greatly increased by the use of the study of polished sections in reflected light. Sulphides also easily replace gangue minerals but the latter

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[1] W. Lindgren, Volume changes in metamorphism, *Jour. Geol.* vol. 26, 1918, pp. 542–554.

[2] Smithsonite often replaces calcite with preservation of structures indicating constant volume. The reaction is supposed to follow the formula $CaCO_3 + ZnSO_4 = CaSO_4 + ZnCO_3$, both $ZnSO_4$ and $CaSO_4$ being water soluble salts. One cubic centimeter of calcite contains 1.192 milligrams CO_2 and 1.518 milligrams CaO while one cubic centimeter of the resulting smithsonite contains 1.514 milligrams CO_2 and 2.787 milligrams ZnO. It is clear then that the principle of equal volumes requires more CO_2 than is available in the calcite. If the process follows the formula, shrinkage of volume will necessarily result.

including sericite, chlorite, calcite, quartz, fluorite and barite very rarely replace sulphides.

Texture of Metasomatic Rocks.—In metasomatism new minerals develop at countless points in the old rock, some growing with crystal form (metacrystic or crystalloblastic series, p. 170) while others grow into irregular grains. Each new grain may be called a metasome, each new crystal a metacryst (pseudophenocryst).[1] The resulting textures will be holocrystalline; the new minerals frequently contain inclusions of the old (sieve texture) and if the replacement is incomplete, as often is the case, enough of the old texture may be preserved to indicate the original rock (relict texture). It is characteristic of some replacements that even if the process has been carried to completion the original texture may be preserved as in silicified oölitic limestone and in silicified dolomites (Fig. 56). In many cases, however, the original texture is wholly destroyed.

Fig. 52.—Replacement veinlets of galena (white) in cryptocrystalline quartz (dark gray) with vugs (black). Tintic, Utah. Magnified 11 diameters.

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[1] This term was first introduced by A. C. Lane, *Bull.*, Geol. Soc. Am., vol. 14, 1903, p. 369.

Grubenmann and Becke use the terms xenoblast and idioblast. N. Grubenmann, Die kristallinen schiefer, Berlin, 1910, p. 91.

FIG. 53.—Feathery geocronite ($5PbS \cdot Sb_2S_3$) (white) replacing galena. Tintic, Utah. Magnified 24 diameters.

FIG. 54.—Galena (light gray) replaced by pearceite ($9Ag_2S \cdot As_2S_3$) (dark gray), in cryptocrystalline quartz. Tintic, Utah. Magnified 227 diameters.

The structure of a rock may be faithfully preserved even when metasomatic action has destroyed its texture. Such preserved structures are, for instance, stratification, joints, breccias, folds and vesicules in lavas. Preservation of texture of limestone which has been completely replaced by sulphides are mentioned by S. F. Emmons[1] and J. M. Boutwell.[2]

It is held by some petrographers that metamorphic rocks show no recognizable succession in order of crystallization but this is certainly not always correct. In many replacements gangue minerals like quartz and barite may crystallize first, while pyrite comes next and other sulphides later.

FIG. 55.—Same replacement magnified 690 diameters. Note that earlier barite plates (black) are not replaced by galena but by the later pearceite.

Irving[3] has pointed out that in some cases replacement begins from a great number of points in the rock where metasomes or metacrysts may develop (Fig. 57) and by continuation of the same process (Fig. 58) the remainder of the rock is finally replaced; the contact is then indefinite. In other cases the complete change occurs rapidly, advancing like a wave over the

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[1] S. F. Emmons, *Trans.*, Am. Inst. Min. Eng., vol. 23, 1893, p. 602.

[2] J. M. Boutwell, *Prof. Paper* 38, U. S. Geol. Survey, 1905, p. 193.

[3] J. D. Irving, *Jour.* Canadian Min. Inst., vol. 14, 1911, pp. 395–471.

country rock; the contacts are then sharp and the process probably consisted in replacement of the original rock by colloid silica. The replaced rock is usually dense and compact; in places, however, drusy cavities occur in it.

Under the influence of the same solution different results may be produced in different rocks. Limestone may be silicified while diorite may be transformed to sericite.

Replacements at High Temperature.—Complete recrystallization, development of silicate minerals with little or no water, and coarse texture are typical of deposits formed by replacement at high temperatures. Mineralizers like fluorine, boron or phosphorus are frequently introduced.

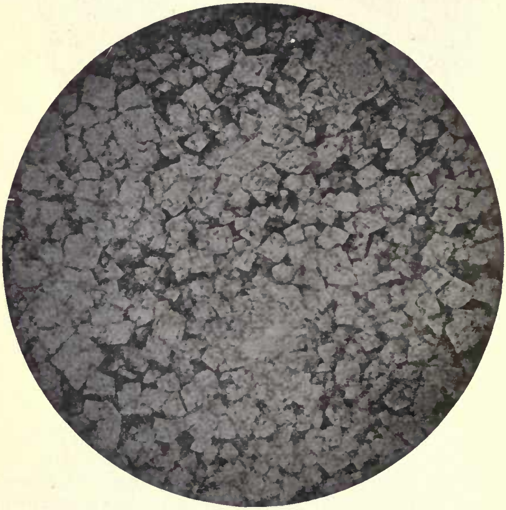

Fig. 56.—Thin section of dolomite completely silicified, but retaining texture and crystal form. *After J. D. Irving, U. S. Geol. Survey.*

The best examples of such textures are found in the replacement of limestones in contact-metamorphic deposits (Figs. 250 and 251) resulting in coarse aggregates of metacrysts of andradite garnet with metasomes of quartz, calcite, epidote and pyroxene. The limestone may be recrystallized in part to coarse calcite. Magnetite and sulphides develop in large grains.

Adjoining tin-bearing veins the rocks are recrystallized to greisen, consisting of coarse metasomes of muscovite, topaz, quartz, tourmaline, fluorite and cassiterite (Fig. 226). Calcareous rocks or greenstones containing much lime are recrystallized to aggregates of axinite, actinolite, garnet, etc.

Replacement by apatite (containing phosphorus), scapolite (containing chlorine) and pyroxene occur adjoining certain high temperature veins. Along many deep-seated gold quartz vein much albite, biotite, and zoisite develop in the wall rock.

Replacements at Intermediate Temperature.—Replacements at more moderate temperature are likely to result in fine-grained textures, and hydrous silicates like chlorite and sericite are abundant (Fig. 61). There are exceptions to this where barite or fluorite replace limestone for both of these minerals easily develop as perfect metacrysts (Figs. 62 and 64). Silicification of limestone, argillaceous shale and rhyolite is a very common process taking place frequently with preservation of texture. The quartz will usually be fine-grained. Silicified limestones are called jasperoids (Figs. 57 and 58).

Limestone may be replaced by massive sulphides (Fig. 63). Alteration by hot waters of granular and porphyritic igneous rocks as well as of schists of similar composition results in sericitization of the femic and salic minerals, sometimes also of the quartz, with development of fibrous aggregates. Pyrite, secondary quartz, rutile, albite and adularia are sometimes found in these rocks. In some classes of such metasomatic rocks carbonates of calcium, magnesium and iron also occur.

Serpentine is altered near some gold quartz veins to coarse aggregates of ankerite, quartz and mariposite (chromiferous sericite).

Quartzite and quartzitic slates in some lead deposits may be extensively replaced by siderite (Fig. 60).

In deposits which have been formed by hot waters near the surface where the rocks are permeable the incipient alteration of igneous rocks is often wide spread with alteration of the femic minerals to chlorite, calcite or epidote (propylitization).

Replacement of effusive rocks by alunite, pyrite and kaolinite is characteristic of some deposits formed near the surface.

Replacement at Low Temperature.—Under the influence of cool solutions the intensity of replacement is diminished. The minerals formed are strongly hydrated, the texture fine-grained.

FIG. 57.—Incipient silicification of limestone. Aspen, Colo. White areas represent quartz crystals with small inclusions of limestone. Magnified 30 diameters.

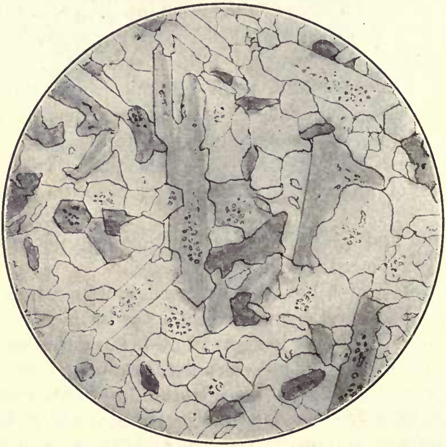

FIG. 58.—Silicified limestone ("jasperoid"). Aspen, Colo. Crossed nicols All quartz. Small inclusions of calcite in some of the grains. Magnified 30 diameters.

In igneous rocks chlorite, quartz and kaolin, possibly also sericite, may form by the action of cool weak solutions. Replacement by sulphides such as pyrite, galena and zincblende may take place Limestone may be silicified to fine-grained jasperoids.

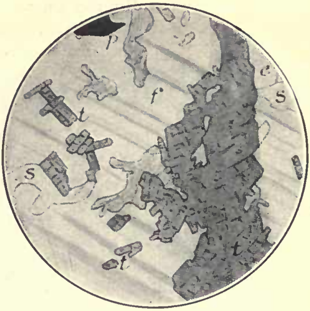

FIG. 59.—Replacement veinlet of tourmaline in fresh andesine grain. Keystone mine, Meadow Lake, Nevada County, Cal. *t*, Tourmaline; *f*, andesine; *e*, epidote; *s*, sericite. Magnified 50 diameters.

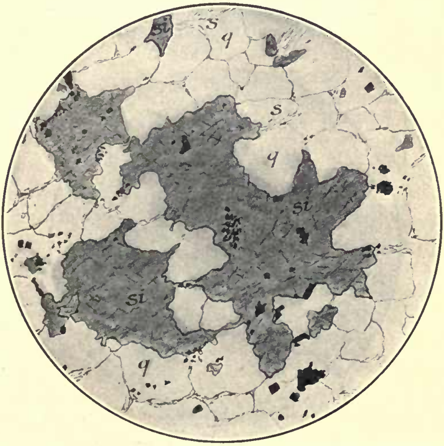

FIG. 60.—Siderite with pyrite and galena, replacing quartzite. Helena and Frisco mine, Coeur d'Alene, Idaho. *q*, Quartz grains; *s*, sericite; *si*, siderite; black, galena and pyrite. Magnified 100 diameters.

To a limited extent sulphides may replace other sulphides. Chalcocite for instance replaces pyrite, chalcopyrite and bornite.

In acid descending waters kaolin replaces sericite and other silicates.

FIG. 61.—Andesine crystal in granodiorite, replaced by sericite and calcite. Pinetree vein, Ophir, Placer County, Cal. *q*, Quartz; *m*, muscovite; *c*, calcite; *s*, sericite. Magnified 80 diameters.

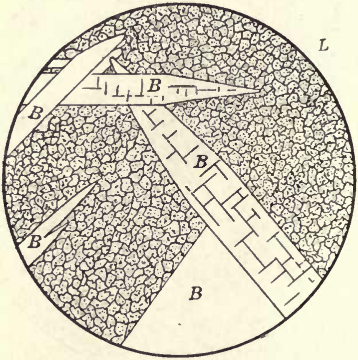

FIG. 62.—Barite (*B*), replacing gray, fine-grained limestone (*L*), Ouray, Colo. *After J. D. Irving, U. S. Geol. Survey.*

Heated alkaline waters are not believed to be capable of developing kaolin from the aluminum silicates of the rocks; alkaline silicates like sericite will result. On the other hand the ordinary dilute ground waters will develop kaolin in the rocks.

In other words kaolin is confined to the uppermost metamorphic zone and rarely ventures far below the zone of weathering.[1]

FIG. 63.—Galena, replacing crystalline dolomite. Elkhorn mine, Montana. *g*, Galena; *p*, pyrite; *c*, calcite grains of limestone; *q*, secondary quartz. Magnified 15 diameters.

FIG. 64.—Fluorite replacing limestone. Florence mine, Judith Mountains, Montana. *f*, Fluorite: *l*, limestone; *q*, secondary quartz. Magnified 7 diameters.

Criteria of Replacement.—F. Posepny first established replacement as a mode of origin of mineral deposits. Shortly after-

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[1] W. Lindgren, The origin of kaolin, *Econ. Geol.*, vol. 10, 1915, pp. 89–93.

ward S. F. Emmons[1] demonstrated it to be a common mode of origin and illustrated it by the description of many ore-bodies in Colorado and elsewhere. About 1900 W. Lindgren described the principal modes of metasomatism.[2] In 1911 J. D. Irving[3] published a paper of great value in which the criteria of replacement ore-bodies were summarized.

Some of these criteria in favor of replacement have already been mentioned but they may be briefly recalled here:

1. Form of ore-body, more or less irregular. Gradually fading limits. Not always conclusive.
2. Presence of unsupported residual rock masses. Sometimes the orientation of bedding may be proved parallel with the surrounding rocks.
3. Absence of crustification. A *banding* may be observed in places due to preservation of bedding or shearing planes.
4. Absence of concave contacts; in limestone, for instance, solution of cavities tends to produce flat concave depressions; a filled cave would show this whereas replacement proceeds with convex outlines toward the unaltered rock.
5. Preservation of textures and structures of original rock. The last-named criterion is the most conclusive.

The criteria for the determination of replacement are sometimes difficult to establish; many mistakes have been made along this line. Replacement veinlets crossing the older minerals and dependence of the replacing mineral on minute fissures and cracks constitute good evidence. The projecting of crystals of one mineral into another is not always a safe proof of replacement. The apparent host may possibly be a later mineral molded about the crystals. In many cases adjoining minerals may have developed practically simultaneously. A peculiar type of replacement results in pseudo-eutectic texture simulating an intergrowth (Figs. 54 and 55).

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[1] S. F. Emmons, The genesis of certain ore deposits, *Trans.*, Am. Inst. Min. Eng., vol. 15, 1887, pp. 125–147.

S. F. Emmons, Structural relations of ore deposits, *idem*, vol. 16, 1888, pp. 804–839.

S. F. Emmons, On the origin of fissure veins, *Proc.*, Colorado Sci. Soc., vol. 2, 1888, pp. 189–208.

[2] W. Lindgren, Metasomatic processes in fissure veins, *Trans.*, Am. Inst. Min. Eng., vol. 30, 1901, pp. 578–692.

[3] J. D. Irving, Some features of replacement ore-bodies and the criteria by means of which they may be recognized, *Jour.* Canadian Min. Inst., vol. 14, 1911, pp. 395–471; *Econ. Geol.*, vol. 6, 1911, pp. 527–561.

Rôle of Colloids in Filling and Replacement.—It is well known that colloid deposits, for instance, of silica, iron hydroxide and aluminum hydroxide play an important part in mineral deposits formed at or near the surface. Colloid minerals are also often deposited during the oxidation of ore deposits. In the discussion in this chapter the colloids have not thus far been considered.

There is, however, an increasing mass of evidence that colloid silica or silica gel is of considerable importance in the origin of deposits formed relatively near the surface by ascending waters. Some of the quartz filling in such veins is extremely fine-grained and bears evidence of having been deposited as a stiff jelly which soon afterward was crystallized in chalcedonic or cryptocrystalline form.[1] Clear evidence of this is seen in some filled veins from the Tintic district, Utah,[2] where the original delicate banding by deposition is still seen though the substance is now microcrystalline quartz.

There is also good evidence presented in the last-named paper on the Tintic district to show that in some deposits formed at moderate depths and not very high temperatures, limestone and dolomite may be replaced by silica gel which afterward crystallized to chalcedony. This type of replacement appears to be characterized by sharp contacts with the unaltered rock; it does not proceed from crystal nuclei of quartz starting at numerous points but advances like wave and stops with sharp contacts (see p. 173). Later metalliferous solutions penetrated this gel and deposited sulphides in it. Sometimes a banding has been produced which strongly recalls the so-called Liesegang rings[3] in artificial gels and indicate a sort of rhythmical precipitation of sulphides.

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[1] W. Lindgren, Geology and mineral deposits of the National District, Nevada, *Bull.* 601, U. S. Geol. Survey, 1915.

[2] W. Lindgren, Processes of mineralization and enrichment in the Tintic mining district, *Econ. Geol.*, vol. 10, 1915, pp. 225–240.

[3] R. E. Liesegang, Geologische Diffusionen, 1913, p. 180. Reviewed by A. Knopf in *Econ. Geol.*, vol. 8, 1913, p. 803.

CHAPTER XIII

ORE-SHOOTS[1]

Form of Primary Ore-shoots.—Commercial ore or mineral does not ordinarily occupy the whole volume of a deposit. The ore is in most cases surrounded by material of poorer grades, sometimes fading into the country rock, or again sharply separated from it. In replacement deposits the disseminated grains of galena, for instance, or zinc blende, may gradually become so few that the mass can no longer be treated with profit. In veins, only certain parts of the sheet-like body can be extracted, while the remainder of the vein material may consist of gangue minerals only, or of clayey attrition masses or breccias.

Those parts of a deposit in which the valuable minerals are so concentrated that their utilization becomes possible are called ore-shoots. Their occurrence and form are exceedingly variable, and it is often most difficult to ascertain the causes which have guided their development. A full discussion of this subject is scarcely possibly here, for it involves the whole question of genesis of mineral deposits.

In deposits of sedimentary origin the ore-shoots have, of course, the general tabular form, but admixture with gangue materials or valueless matter may so dilute the ore that only certain parts of the body can be extracted. Various assort-

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[1] C. R. Van Hise, Some principles controlling the deposition of ores, *Trans.*, Am. Inst. Min. Eng., vol. 30, 1900, pp. 27–177.

T. A. Rickard, The formation of bonanzas in gold veins, *Trans.*, Am. Inst. Min. Eng., vol. 31, 1902, pp. 198–220.

The localization of values in ore-bodies, etc. Discussion by J. D. Irving, F. C. Smith, Reno Sales, F. L. Ransome, H. V. Winchell, H. Sjögren, and W. Lindgren, *Econ. Geol.*, vol. 3, 1908, pp. 143–154, 224–229, 326–330, 331–336, 425–427, 637–642; vol. 4, 1909, pp. 56–61.

C. W. Purington, Ore horizons in the San Juan Mountains, *Econ. Geol.*, vol. 1, 1905–06, pp. 129–133.

H. C. Hoover, The valuation of gold mines, *Eng. and Min. Jour.*, May 19, 1904.

R. A. F. Penrose, Jr., Some causes of ore-shoots, *Econ. Geol.*, vol. 5, 1910, pp. 97–133.

ments of detritus and complex conditions of precipitation from waters of seas, lakes, and rivers have influenced the concentration of the richer ore masses. In addition, alterations by meteoric waters are common; in the case of phosphate deposits and beds of siderite they have resulted in enrichment.

In deposits of igneous origin the general form of the deposit is also that of the ore-shoots. In some deposits, such as the magnetite deposits of northern Sweden and the dike-like deposits of ilmenite at Iron Mountain, Wyoming, there is practically no waste material and the whole igneous body constitutes ore.

More commonly the irregular lenticular or tabular masses of igneous rocks in which ore minerals have developed by magmatic segregation (for instance, gabbro containing chalcopyrite) have nuclei of richer material gradually fading into more normal rock.

In the epigenetic deposits the outlines of the ore-shoots are exceedingly variable. In those deposits which are formed by replacement this is particularly true, and few rules can be laid down for their occurrence; the form is determined by the fissures giving access to the solutions, by the presence of impermeable rocks, and by the varying susceptibility to replacement of the original rocks.

Most attention has been given to the shoots in fissure veins. Although the ore in the main follows the fissure and therefore has a tabular or sheet-like form, it rarely occupies the whole space along this fissure, but is concentrated in bodies of varying size, shape, and continuity; smaller bodies are known as bunches, pockets, or kidneys; in gold-quartz veins these may be exceedingly rich. Narrow ore-shoots, greatly elongated in the vertical direction, whether occurring in fissure veins or independently of them (for instance, in volcanic necks), are called *chimneys*, *pipes*, or *necks* (Fig. 68).

Ore-shoots may be entirely irregular, but commonly have a more or less well-defined columnar, steeply pitching shape, best shown in projection upon the plane of the vein. Fig. 65 shows the terminology proposed[1] for various dimensions of an ore-shoot in a vein. The pitch length, or axial length, is the distance between the two extreme ends of the shoot. The pitch is the angle

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[1] W. Lindgren and F. L. Ransome, *Prof. Paper* 54, U. S. Geol. Survey, 1906, p. 206.

which the pitch length makes with the strike of the vein, and is measured on the plane of the vein. The stope length is the horizontal length of the ore-shoot on any particular level. The thickness or width is measured perpendicularly to the plane of the vein. The breadth of the ore-shoot is the stope length, multiplied by the sine of the pitch.

Fig. 66 shows the ore-shoots of a gold-quartz vein at Nevada City, California. Flat-dipping shoots are not so common. Fig. 67 shows an excellent example of a flat shoot in the celebrated Eureka-Idaho vein at Grass Valley, California.

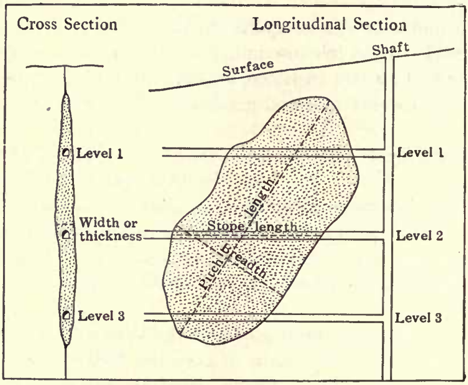

Fig. 65.—Diagram illustrating the terms used to describe the dimensions of ore-shoots. *After W. Lindgren and F. L. Ransome, U. S. Geol. Survey.*

In parallel veins the shoots are often, roughly speaking, coextensive. Sometimes the shoots in a series of parallel veins persistently recur where the veins cross a certain stratum or dike, as, for instance, where the gold-quartz veins of Gympie, Queensland, intersect certain carbonaceous strata, or as at Thames, New Zealand, where the veins intersect certain softened and altered andesites. Many shoots follow intersections of veins or of veins with fissures.

Shoots, however large, do not continue indefinitely, but end in depth, usually with gradual deterioration. Small masses or kidneys are likly to be found below the termination of a large ore-shoot. Exploration may find another shoot below the first, either on the same fissure or imbricating on a parallel

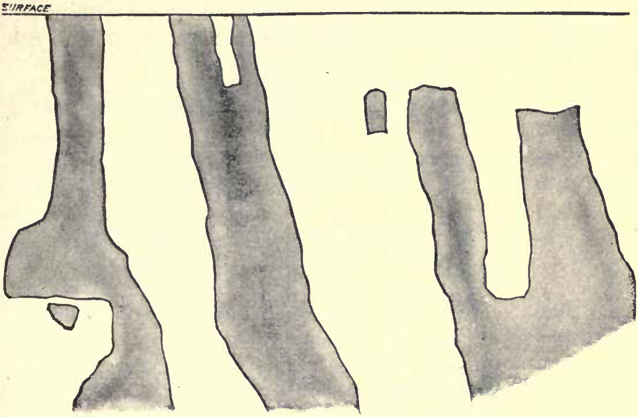

FIG. 66.—Ore-shoots of veins at Nevada City, California.

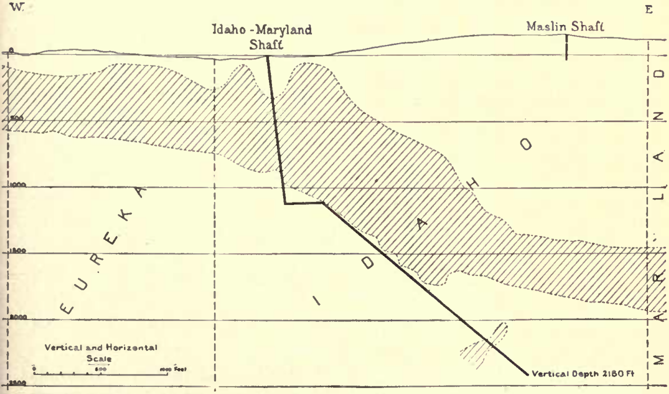

FIG. 67.—Approximate outline of the Eureka-Idaho ore-shoot, Grass Valley, California, in projection on the plane of the vein.

vein. When great depth is attained the grade of the ore usually decreases in the deeper levels, but this rule is not without exceptions. Many shoots are lenticular, that is, they contain a rich nucleus, outward from which the ore gradually decreases in tenor. H. C. Hoover, from an examination of 70 mines, concluded

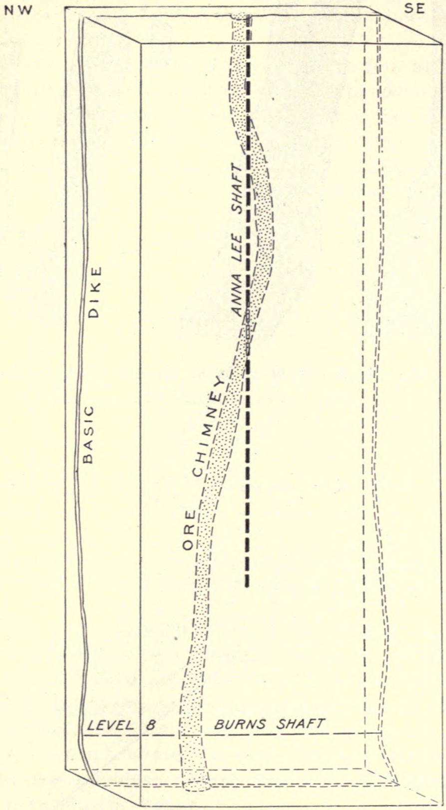

FIG. 68.—Stereogram of Anna Lee ore chimney, Cripple Creek, Colorado. Shoot probably determined by intersection of the basic dike with a fissure. *After V. G. Hills.*

that ore-shoots are generally lenticular and that the probable minimum extension of an ore-shoot below any given level would be a factor of not less than a radius of one-half of its breadth.

At Cripple Creek Lindgren and Ransome found that the shoots which begin distinctly below the surface have a marked elongated form, the ratio between pitch length and breadth varying from 1½ : 1 to 5 : 1.

Primary ore-shoots rarely continue for more than 2,000 feet along the strike, or for more than 2,000 feet along the pitch length.

In a given district the pitch of the ore-shoot is often predominantly in one direction; thus at Nevada City and Grass Valley the shoots pitch to the right of an observer who looks down the dip of the vein. In another district the opposite may be true. In some places the tenor varies directly, in others inversely with the swelling of the vein. According to a rule often quoted, the shoots follow the directions of the striations on the vein walls, but this again by no means has universal application.

Shoots of Successive Mineralizations.—While in some veins the whole width consists of uniform ore, it is exceedingly common, especially in thick veins, to find that there are certain streaks which are far richer than the rest. They may follow foot-wall or hanging-wall, or the center of the vein, or may switch from one side to another. Such phenomena indicate re-opening of the vein or brecciation after the first period of vein-filling and enrichment.

Superficial or Secondary Shoots.—Descending surface waters decompose and often enrich the upper part of veins or other deposits. Such enriched superficial portions of an ore deposit are dependent upon the ground-water level and, when projected upon the plane of the vein, follow the surface of the ground and terminate below along an irregular and jagged line. Oxidized ores, as well as sulphides due to enrichment, are found in them, usually at different levels. The surface shoots are in fact characterized by horizontal extension, in contradistinction to the predominance of the vertical direction in the primary shoots. The mineralogical characteristics of superficial shoots will be discussed in detail in a later chapter. Their tendency is to spread along the strike of the vein, often also out into the wall rock. Thus pay ore may be found for a long distance along the trend of the vein and its appearance will be that of the oxidized croppings of a long primary shoot, when in fact deeper explorations may prove the existence of only a few narrow primary ore-bodies underneath the continuous surface ore. Sometimes, as in Calico, San Bernardino County, California, and numerous other places, oxidized silver ores will be found in croppings along a vein which are simply concentrations of a primary vein filling that contains no workable shoots. To this class belong also the horizontal or flat shoots of secondary copper sulphides (chalcocite and covellite) formed by descending solutions in copper deposits at or near the water level. The primary material may or may not constitute commercial ore. If spread over wide mineralized areas such shoots are often called *chalcocite blankets.*

Lateral spreading is often characteristic of shoots of oxidized ores. Descending metal solutions may wander out in the country rock and here form new bodies.

Causes of Primary Ore-Shoots.—Ore-shoots are due to the abundant precipitation of valuable minerals from their solutions. The causes are in part chemical and in part mechanical:

1. Decrease of pressure and temperature.
2. Favorable chemical character of wall rock.
3. Favorable physical character of wall rock.
4. Intersections.

Decrease of Pressure and Temperature.—The fundamental reason for the occurrence of ores in veins and allied epigenetic deposits in the upper crust is probably that the metals were in solution in hot waters which were ascending and gradually encountered conditions favorable for precipitation. First among these conditions is decreasing temperature. If this is true the deposits should gradually become poorer or barren in depth.[1] In a general way this is doubtless true, but for many substances the vertical space through which deposition can take place is very

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[1] T. A. Rickard, Persistence of ore in depth, *Trans.* Inst. Min. and Met., vol. 24, 1915, pp. 3–46, with discussion.

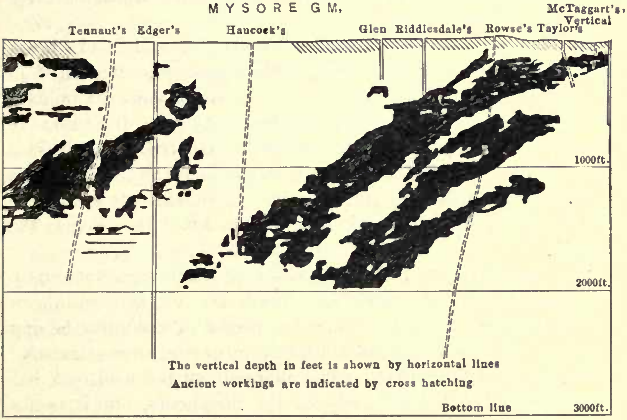
W. Lindgren, Ore deposition and deep mining, *Econ. Geol.*, vol. 1, 1905, pp. 34–46.

F. L. Garrison, Decrease of value in ore-shoots with depth, *Trans.* Canadian Min. Inst., vol. 15, 1912, pp. 192–209.

J. F. Kemp, The influence of depth on the character of metalliferous deposits, Compte Rendu, 12*e* Session, Canada, Congrès géologique internat., 1914, pp. 253–260.

Malcolm Maclaren, *Idem*, pp. 295–304.

large. We know that gold-bearing quartz was deposited in California over a vertical distance of 4,000 feet, while in southeastern Alaska and at Bendigo, Australia, the interval is not less than 5,000 feet. This deposition took place at considerable depth below the surface, probably several thousand feet below it, and as it is known that gold-bearing quartz may also be deposited within the upper zone, we have thus a total vertical range of at least 9,000 feet. At the lowest levels at the places mentioned the ore is of low grade, but in Alaska at least there is a large quantity available. The richest ore was doubtless deposited close to the surface, where we find the bonanzas of the Tertiary gold and silver veins; but below this bonanza zone the decrease in tenor of the ore is very slow and rich shoots and pockets may be found at great depth below the original surface. The most persistent gold-bearing ore shoots known are those in veins formed at intermediate or high temperatures. Such are, for instance, the North Star vein at Grass Valley, California, which with very slight impoverishment has been followed for 6,400 feet on a dip of 20° (p. 569). The Kolar veins in India have been mined to a vertical depth of 4,000 feet in shoots of considerable regularity (Fig. 69), and with little change in tenor of ore.



The vertical depth in feet is shown by horizontal lines
Ancient workings are indicated by cross hatching

FIG. 69.—Pitching ore shoots in gold quartz veins, Kolar, India. *After T. A. Rickard.*

The most persistent ore body known is that of Morro Velho mine in Brazil, where a pitching ore shoot has been worked to a vertical depth of 6,200 feet and a pitch length of 9,000 feet (Fig. 236). For copper ores the vertical range of deposition is likewise great, though unlike gold and silver they seem to be deposited in greatest quantity at lower levels and high temperatures. Lead, on the other hand, appears to be precipitated nearer the surface and at lower temperatures; while zinc in this respect stands between copper and lead.

The relations set forth explain why so little decisive evidence of vertical succession in deposition is available from observations at any one mine.

In the Cornwall veins tin and tungsten prevail in the lower levels in granitic country rock, while copper was deposited in the cooler region of the slates covering the granite batholiths; the lead ores are found some distance away from the intrusive granite. In many lead mines it has been noted that within a distance of 700 to 3,000 feet from the surface the lead minerals give way to pyrite and zinc blende. In quicksilver mines the ore often becomes impoverished within 1,000 feet below the surface.

The dependence of the deposition of various metals upon temperature and therefore also upon the vertical and horizontal distance from the place of origin of the mineralizing solutions has been emphasized lately by several investigators.[1]

Character of Wall Rock.—The character of the wall rock has sometimes a decided influence on the ore-shoots, but it is not always easy to decide whether it is due to chemical or mechanical causes. In replacement deposits limestone and lime shale are usually favorable, but in the Coeur d'Alene district of lead-bearing veins a quartzitic schist is the rock best adapted for replacement by siderite and galena. At Freiberg, Saxony, the gray gneiss is the favorable rock, while the veins split or become unproductive in the red gneiss or in the mica schists.

Carbonaceous rocks are believed to influence deposition favorably by their reducing action; the gold-quartz shoots of Gympie,

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[1] J. E. Spurr, A theory of ore deposition, *Econ. Geol.*, vol. 2, 1907, p. 790.

L. De Launay, La metallogénie de l'Italie, Congrès géologique internat., Mexique, vol. 1, 1906, p. 571. Also in Gîtes Minéraux, vol. 1, Paris, 1913.

W. Lindgren, Processes of mineralization and enrichment in the Tintic mining district, *Econ. Geol.* vol. 10, 1915, p. 228.

Queensland, are often quoted, as well as the supposedly carbonaceous "indicator" at Ballarat, Victoria. The well-known replacement of fossil wood by chalcocite in a certain class of copper deposits may be added to these examples, as well as the supposed influence of certain oil shales on the deposition of lead ores in Wisconsin. The importance of precipitation by carbonaceous material has been overestimated, but in many cases the hydrocarbons have certainly favorably influenced the deposition of ores.[1]

Rocks containing pyrite or other sulphides often enrich traversing veins. Examples of this are known from Kongsberg, Norway, where the silver veins are productive when crossing certain schists with disseminated sulphides. At Ophir, California, gold-quartz veins are enriched when crossing "iron belts" of pyritic amphibolites.

Where a vein cuts through a thick series of sedimentary rocks it often widens and contains rich ore in the limestones, while poor or barren in shale or sandstone. Similarly, where a thick series of igneous rocks, as in the San Juan region, Colorado, is intersected by veins ore horizons will develop in rocks which by their physical and chemical character are most favorable to continuous fissures or to replacement.

Rhyolites are generally unfavorable because fissures often tend to splitting in such rocks; tuffs likewise because the solutions tend to disperse through great masses of rock.

On the other hand, rocks like andesites and latites are usually favorable. Purington (*op. cit.*) has shown that in the San Juan Mountains the andesitic breccias which contain abundant ferromagnesian silicates are most favorable to ore deposition.

Impermeable Barriers.—The conditions outlined above would tend to produce more or less horizontal ore-bodies. Such ore-bodies are most conspicuous where impervious rocks interpose barriers to the solutions. The occurrence of ore in horizontal extension *below* such barriers is in fact one of the best indications that the solutions have been ascending in the main.

Fig. 70 shows the occurrence of oxidized silver ores below the Devonian shale at Chloride Flat (Silver City), New Mexico, and similar occurrences are not uncommon in other mining districts of New Mexico. The blanket veins of Rico, Colorado (Fig. 71),

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[1] W. P. Jenney, The chemistry of ore deposition. *Trans.*, Am. Inst. Min. Eng., vol. 33, 1903, pp. 445–498.

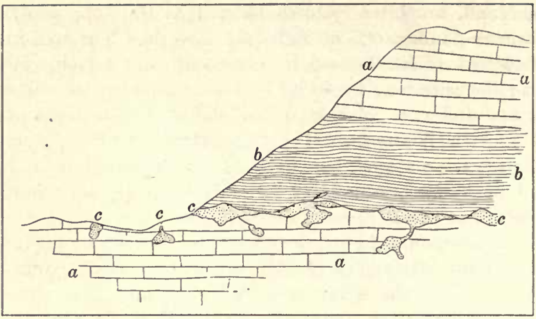

FIG. 70.—Sketch section showing occurrence of ore-shoots in limestone at contact of overlying Devonian shale at the Bremen mine near Silver City, New Mexico. *a*, Limestone; *b*, shale; *c*, ore. *After R. A. F. Penrose, Jr.*

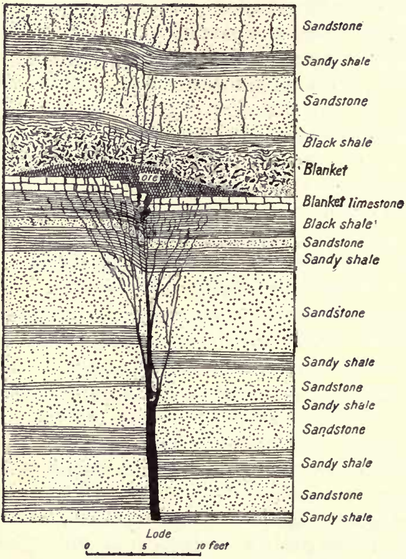

FIG. 71.—Diagrammatic section across a lode, and ore-body formed beneath an impervious stratum (blanket) of black shale, Rico, Colorado. *After F. L. Ransome, U. S. Geol. Survey.*

present another good illustration of this principle, as do also the ores of the American Nettie mine near Ouray, Colorado, and the siliceous gold ores replacing dolomite in the Black Hills.

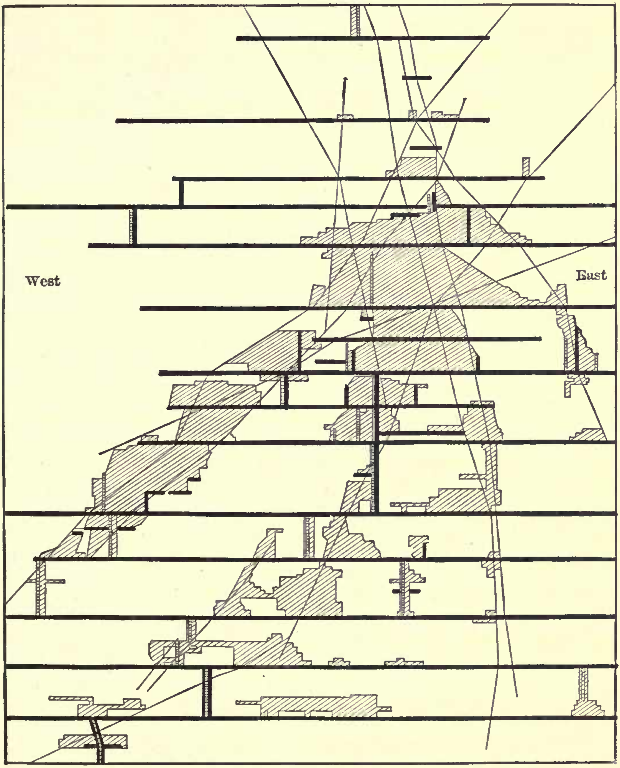

Fig. 72.—Longitudinal section along the Neu Hoffnung vein, Freiberg, Germany, showing ore-shoots along intersection with several other veins. *After R. Beck.*

The impermeable stratum is not necessarily shale; it may be a gouge in a fissure, or a sheet of volcanic rock which, for some reason, the fissures failed to penetrate. The same principle of impermeable barriers serves to explain why the vein material is often confined between the clay seams of hanging and foot wall without entering the adjacent country rock by replacement.

Where one fissure is faulted by another, deposition may occur because the circulation becomes impeded at the fault. It is not entirely clear why deposition of rich ores should take place when the solutions are impeded and partial stagnation follows, but the conditions observed bear sufficient testimony to the fact.

Where the solutions have moved downward, as in the concentration of hematite ore from poorer "iron formations," it is often observed that ores occur on *impervious basements* and in *troughs* caused by shales, clayey fissures, or dikes.

Intersections.—Enrichment and ore-shoots along intersections of two veins or of a vein and a fissure are very common phenomena, well exemplified at Freiberg, Saxony (Fig. 72), and at Cripple Creek, Colorado. Van Hise attributes the shoots at such intersections to the mingling of two solutions and consequent precipitation of some constituents. In part they may be due to the shattering of the rocks at the intersection, and Penrose notes that shoots are more likely to occur where the intersection takes place at acute angles, forming wedge-shaped blocks that are easily broken along their edges.

Though enrichment at intersections is common it is by no means a universal rule, and indeed sometimes a vein is impoverished at the intersection with a barren fissure.

The occurrence of the large shoots such as those in the gold-quartz veins of California, at Cripple Creek, and in the Coeur d'Alene lead mines cannot be fully explained by intersections or by the influence of the wall rock.

Such shoots are generally considered as the result of decrease in temperature of ascending solutions in channels of circulation.

CHAPTER XIV

THE CLASSIFICATION OF MINERAL DEPOSITS

Classification by Form and Substance.—A genetic classification of deposits of useful minerals is really equivalent to the classification of "geological bodies" as definied in Chapter I and is therefore naturally beset with all the difficulties connected with an imperfect knowledge of geological processes. The early attempts in the way of systematic treatment, however, avoided this troublesome path by the simple expedient of classifying by substance or uses, or by form. These schemes are followed in many text-books, even among those of recent date; undoubtedly they have some advantages, especially for the miner, the industrial chemist, or the metallurgist, who are principally interested in the form of the deposit or in the study of ores of certain metals.

By substance and uses mineral deposits may be classified as follows:

1. Structural materials	Stone, glass sand, cement rock, clay, asphaltum.
2. Fuels	Coal, petroleum, natural gas, peat.
3. Abrasives	Corundum, garnet.
4. Fertilizers	Potash salts, phosphates, green-sands.
5. Precious stones	Diamond, opal, tourmaline.
6. Various industrial uses	Graphite, barytes, borax, asbestos. sulphur.
7. Metallic ores	Iron ores, copper ores, gold and silver ores, tin ores, aluminum ores, etc.

However convenient, it is evident that this classification cannot lead to a thorough appreciation of the manifold processes by which mineral deposits are formed in nature.

The first attempts at a classification of the deposits themselves were made by the miners and thus the early and not yet entirely abandoned schemes refer to the form of the geological bodies. But form is closely connected with genesis and even in one of the earliest classifications on this basis, that of Bernhard von Cotta,[1]

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[1] Die Lehre von den Lagerstätten, Freiberg, 1859.

the difficulty of avoiding genetic conceptions is felt in his definition of a vein as a "filled fissure." He divides ore deposits as follows:

- I. Regular deposits.
  - *A*. Beds.
  - *B*. Veins.
    - *a*. Ordinary fissure veins (true fissure veins).
    - *b*. Bedded veins.
    - *c*. Contact veins.
    - *d*. Lenticular veins.
- II. Irregular deposits.
  - *C*. Stocks. (Irregular masses with distinct limits.)
    - *a*. Recumbent.
    - *b*. Vertical.
  - *D*. Impregnations. (Irregular masses, fading into country rock.)

With variations this plan of classification is followed in many of the older text-books. Not unlike it is a classification by J. A. Phillips in his treatise on ore deposits, revised in 1896 by H. Louis.

Lately James Park has adopted the same plan with some modifications in a useful and practical text-book on mining geology.[1] His classification is as follows:

- I. Superficial deposits.
  - *a*. Fragmentary.
  - *b*. Massive.
- II. Stratified deposits.
  - *a*. Constituting beds.
  - *b*. Disseminated through a bed.
- III. Unstratified deposits.
  - *a*. Deposits of volcanic origin.
  - *b*. Stockwork deposits.
  - *c*. Contact or replacement deposits.
  - *d*. Fahlbands.
  - *e*. Impregnations.
  - *f*. Segregated veins.
  - *g*. Gash veins.
  - *h*. True fissure veins.

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[1] A text-book of mining geology, London, 1907, p. 219.

Park states that this classification is only an empirical arrangement to facilitate the study of ore deposits, and a provisional classification by origin also is given.

L. De Launay[1] arranges the deposits according to the principal elements contained. This logical, though not genetic, plan has been followed in part in the index appended to this book.

Genetic Classifications.—A genetic classification is the most desirable both theoretically and practically. In exploring and exploiting ore deposits, the miner is almost forced to form an idea of its origin in order to follow up the ore-bodies to best advantage. Von Groddeck and Stelzner were really the first mining geologists who appreciated and applied the genetic principle in classification.[2] Of course, the time was hardly ripe for its introduction until the conceptions of genesis had crystallized into fairly definite form. Stelzner remarks, with good reason, that it is only by standing upon the ground of a genetic theory that the miner finds courage to sink deep shafts or drive long tunnels.

We are still in doubt as to the true mode of origin for many deposits. But, as von Groddeck and Stelzner have pointed out, this applies to any classification and this very uncertainty is a stimulus to further investigations.

The different classifications proposed will not be given here in detail. An excellent account is found in Kemp's "Ore deposits of the United States and Canada," Appendix I. Von Groddeck and Stelzner, Posepny, Wadsworth, Monroe, Kemp, Crosby, Hoefer, Spurr, Van Hise, Weed, and several others have more or less successfully attacked the problem of a consistent genetic classification.

Von Groddeck, followed by Stelzner and Beck, makes the primary distinction whether the useful minerals were originally formed in or with the rock in which they now occur or whether they were introduced into pre-existing rocks. Stelzner called the former syngenetic, the latter epigenetic. (Author's lecture notes, Freiberg, 1881.)

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[1] L. De Launay, Gîtes Minéraux et Métallifères, 3 vols., Paris, 1913.

[2] The former says: "I must confess that I have never been able to understand the satisfaction which many people feel when they are informed that a certain deposit, for instance, is a 'stock.' This information has, on the contrary, always produced in me a feeling of deep dissatisfaction." Quoted in Stelzner and Bergeat, Erzlagerstätten, pt. 1, 1904, p. 10.

J. F. Kemp divides the deposits into (I) those of igneous origin, (II) those precipitated from solutions, and (III) those deposited from suspension, or residues after the decomposition of rocks. Difficulties appear here too, for what are igneous magmas but solutions?

Beck's classification is in part based on that of Stelzner. In the first edition of his hand-book "Die Lehre von den Erzlagerstätten" the syngenetic or epigenetic origin was made the principal basis of classification. In the edition of 1909 this is changed and the deposits are classified as follows, on the basis of the various phases of their genetic history:

1. Magmatic segregations.
2. Contact-metamorphic ore deposits.
3. Fissure veins.
4. Bedded deposits.
5. Stocks.
   Morphologic facies of a single genetic group.
6. Secondary alterations.
7. Sedimentary ore deposits.
8. Detrital deposits.

While this is a decided improvement upon the first classification adopted by Beck, the description of the various deposits shows that many genetically different types are forced into one and the same subdivision.

Weed[1] goes further and gives the origin of the ore-forming solutions. His first class includes igneous deposits, segregated in a magma; his second, igneous emanations, including contact deposits, and tin veins; his third, gas-aqueous or pneumatohydato-genetic deposits formed by magmatic waters mingled with ground waters. His fourth and smallest division includes those mineral masses formed by surface waters. This classification has not been generally accepted because it brings up the admittedly difficult separation of meteoric and magmatic water.

The best genetic classification of mineral deposits would seem to be that according to geological processes. Mineral deposits must have been formed by igneous processes, alteration, cementation, deformation, erosion, or sedimentation. Recognizing this, Van Hise[2] classifies ores as follows: Those produced (1) by proc-

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[1] W. H. Weed, In "Ore deposits," a discussion republished from the *Eng. and Min. Jour.*, New York, 1903, pp. 20–23.

[2] C. R. Van Hise, A treatise on metamorphism, *Mon.* 47, U. S. Geol. Survey, 1904.

esses of sedimentation; (2) by igneous processes; (3) by metamorphic processes, including under this heading practically all veins and allied geological bodies, conceiving them to be deposited by the circulating ground water.

It is probably impossible to produce a classification which will win the approval of all. In the ultimate analysis by far the larger number of mineral deposits have been formed by physicochemical reactions in solutions, whether these were aqueous, igneous, or gaseous. According to this view the only consistent division that can be made is that between deposits formed by mechanical concentration of pre-existing minerals and those formed by reactions in solutions.

A genetic classification should not be confined to a general indication of the relative time of ore deposition—whether at the same time or later than the country rock. Nor should it confine itself to a statement of the agents of ore deposition—whether aqueous, igneous, or gaseous solutions, or whether sedimentary, igneous, or metamorphic processes. The statement of the place of ore deposition—at the surface or below it; in shallow waters or in deep seas—is important but not sufficient.

Some authors have attempted a classification by mode of deposition—whether by replacement or by filling of open cavities—but all such attempts have been failures, for the two processes are so closely associated that separation is impossible.

The genetic classification should ultimately determine the limits of ore deposition in each class by temperature and pressure. Each deposit should be considered as a problem in physical chemistry, and the solution of this problem, with the necessary geological data, will suffice to fix the mode of formation of the deposit.

We are far from having the complete material for such a classification, but we have at least a few starting points. It is necessary to determine, by experiment or by observation in nature, the limits of existence of each mineral species. Some will be found to be "persistent" under widely differing conditions of temperature and pressure—like fluorite, quartz, or gold. For others a far more limited range will be established. By collecting the data of mineral association, sequence of deposition, and stability range of the component parts of the deposit it will be possible to ascertain the conditions prevailing at the time of ore deposition.

An absolutely consistent genetic classification is at present impracticable for its forces the geologist to take a definite stand on problems which, as yet, have not been solved.[1]

Perhaps it is well not to expect too much from physical chemistry, magnificent as its services have been. The complications, even in simple systems, become great when, besides temperature and pressure, concentration, mass action, and time must be considered. In multicomponent systems the difficulty increases enormously. At the same time it is believed that the direction indicated is the only safe one to take in classifying the complex phenomena of ore deposition.

OUTLINE OF PROPOSED CLASSIFICATION

Detrital and Sedimentary Deposits.—In the scheme followed in this book there are two major divisions. The first includes deposits formed by mechanical processes of concentration. This includes the detrital deposits such as placers and quartz sand formed at moderate temperature and pressure.

The second division contains the great majority of mineral deposits which have been produced by chemical processes of concentration. Many important processes, such as those productive of iron ores and phosphates, for instance, take place by interactions of solutions in bodies of surface waters. These processes may be of inorganic origin or they may take place through the medium of living bodies, almost always at moderate temperatures. The products are usually mingled with detrital matter. They may be enriched by secondary processes in the unconsolidated strata or by processes of weathering after their exposure to air.

Another class of deposits is formed in bodies of surface waters by their evaporation and consequent precipitation of the salts dissolved in them; these are frequently termed the "saline residues." Common salt, gypsum, and borates are among the substances found in these deposits.

Concentration of Substances Contained in the Rocks.—Instead of at the surface or in bodies of surface waters the processes of concentration of useful substances may go on in the rocks themselves. We may distinguish two cases: the sub-

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[1] T. Crook, The genetic classification of rocks and ore deposits, *Mineralog. Mag.*, London, vol. 17, 1914, pp. 55–85.

stances were originally contained in the same geological body in which the deposit is found, or they may have been introduced from the outside.

The apparent objection to this basis of subdivision, namely, the difficulty of deciding the source of the mineral or metal, is met in many cases by the knowledge acquired during late years. There may be deposits for which the qestion cannot be decided, but I believe that in the near future we shall in most cases have sufficiently good evidence. No one seriously maintains that the gold in the quartz veins of California, for instance, has been leached from the surrounding country rock, and surely no one denies that the oxidized nickel silicate ores of certain peridotites were originally contained in minute distribution in these rocks.

In the case of substances contained in the geological body itself, the concentration may be effected by (1) rock decay and residual weathering—that is, by oxygenated surface waters; (2) by the ground water of the deeper circulation; (3) by processes of dynamic and regional metamorphism, and (4) by zeolitization of surface lavas.

Residual Weathering.—Rock decay tends to destroy the rocks as units; to break them down, mechanically and chemically, and to re-assort their constituents in new combinations. In the decaying mass certain constituents are concentrated or precipitated; its detritus is swept away and deposited in rivers, lakes, and oceans; its soluble constituents are carried into the larger reservoirs of water and there perhaps precipitated in various forms.

It is true that not quite all the sedimentary deposits are derived from the decaying rocks; the fossil coals are indirectly made from the carbon of the atmosphere; volcanic ashes contribute a share to the sediments; the exhalations of eruptive magmas, as well as ascending waters, contribute some dissolved matter from the lower part of the earth's crust.

Processes of sedimentation and rock decay take place at moderate temperatures and pressures and the new minerals formed are, as a rule, characterized by high hydration. Below 0° C. mineral deposits do not form, except in so far as freezing of water is retarded by rapid motion or dissolved salts. Few of the deposits have been formed at temperatures above 50°, and this only exceptionally during eruption, evaporation in shallow desert lakes, or oxidation of pyritic rocks. The pressure is in general little different from that of the normal atmosphere, but in deposits of deep seas or lakes considerably higher pressures prevailed. This increased pressure, at low temperature, appears to have had little influence on the mineral associations formed.

Deep Circulating Waters.—Under the influence of the ground water of the deeper circulation many ore deposits are formed, concerning some of which there may be room for differing opinions. Copper may be leached from greenstones and the ores of the metal may be deposited in veins in the same rock. Hematite, like that of the Lake Superior region, may be concentrated from the surrounding low-grade "iron formation." Barite, magnesite, and sulphur are other instances.

Regional Metamorphism.—Again, the agency may be metamorphism under stress or regional metamorphism; in such case the change takes place with very little water and it is not considered probable that a great concentration, say of the metals contained, can be effected. Other materials may form, such as slate from shales, or useful minerals like garnets or graphite may develop in the rock. During static metamorphism, temperature and pressure are likely to be somewhat higher than at the surface. Regional metamorphism takes place under heavy pressure and at fairly high temperatures at great depth. It may merge into igneous metamorphism.

Zeolitization.—The processes of zeolitization take place shortly after the consolidation of an igneous rock by the aid of residual magmatic water or of surface water. Under certain circumstances a concentration of metals can be effected by this process, of which the copper deposits of Lake Superior offer an excellent instance.

Introduced Ores not Connected with Igneous Rocks.—Much more common is the case where the valuable minerals have been introduced into the rock from without, and to this class belong the majority of the metal deposits. Deposits of this kind occur along fissures or form replacements along fissures or are found in general where opportunity is offered for vigorous circulation of the depositing waters. For a long time it was held by many that the metallic contents of fissure veins were derived from the surrounding rock, but it is now generally admitted that such a view in most cases is erroneous.

Certain metallic ores occur entirely independent of igneous rocks; the mineral associations in these indicate a deposition at moderate pressure and temperature, the latter probably rarely reaching 100° C. Of this kind are certain lead-zinc deposits in limestone or the copper deposits in sandstone which are so common in various parts of the world. Most geologists agree that such deposits have been formed by surface waters, at moderate depths; and that the metals have been leached from neighboring strata and, after a comparatively short wandering, deposited in fractured rocks in their present resting places. These deposits are generally poor in gold and silver.

Deposits Genetically Connected with Igneous Rocks.—There is also another and larger class which appears only in or near igneous rock and whose epoch of formation usually can be shown to have followed closely after the eruption. This class has been clearly recognized by almost all geologists. There is also general agreement that these deposits have been laid down by heated, ascending waters, although there is no unanimity as to the source of the water or the source of the metal. To some the water and the dissolved metals are simply igneous emanations from a cooling magma; to others the waters are of atmospheric origin and, heated by their passage through still warm igneous rocks, have dissolved the metals contained in them.

Nearly all metal deposits of the American Cordilleran region belong to this division. It is subdivided into several groups, according to the evidence of mineral association and geological relations. The first group includes ores deposited at slight depth below the surface; the temperature is here relatively low, perhaps from 50° to 150° C., and the pressure will scarcely exceed 100 atmospheres. Examples of this group are found in the gold and silver veins, of Tonopah, Nevada, the Cripple Creek gold telluride veins, and the California quicksilver veins.

A second group comprises the deposits formed by hot ascending waters at moderate depths, say from 5,000 feet to 10,000 feet below the surface, at temperatures of perhaps from 150° to 250° C. and correspondingly increased pressure. The present outcrops are exposed by deep erosion and they almost always appear in or close to intrusive bodies. As examples may serve the gold-quartz veins of California and the metasomatic pyritic deposits of Leadville.

A third, deep-seated group includes veins and contact-metamorphic deposits. During the genesis of these the temperature was high, but in most cases below 575° C., the crystallographic inversion point for quartz. The pressure was probably very high. The cassiterite veins, some gold-quartz veins of the Appalachian type, and the tourmaline-copper veins belong in this group, which with great confidence may be ascribed to emanations from magmas. The deposits unquestionably formed by direct igneous emanations are the contact-metamorphic ores appearing in limestone along igneous contacts. They contain oxide ores, such as magnetite and specularite, together with sulphides of copper, zinc and iron, and present an association of other minerals characteristic of contact metamorphism

The emanations from effusive bodies are deposited as sublimates of little economic importance.

Products of Magmatic Differentiation.—The last class is that of the deposits formed by concentration in igneous magmas; of all types these have formed at the highest temperature and pressure. They include oxides or sulphides segregated in the magmas, like the iron ores of Kiruna in northern Sweden, the titanic iron ores of the Adirondacks, or the copper-nickel ores of Sudbury. They also include the pegmatite dikes, which contain many gems and rare metals and which are regarded as segregations from cooling granitic magmas. The pegmatites were formed at comparatively low temperatures—probably from 500° to 800° C.—but during the differentiation of the other deposits mentioned considerably higher temperatures probably prevailed. The pressure must, of course, have been very high.

Metamorphism and Surface Enrichment of Deposits.—In the proposed classification the mineral deposits are supposed to have suffered no change from their original condition. This is of course rarely strictly true, for chemical changes as a rule begin soon after the cessation of the agency which caused the deposition. In sedimentary beds this is particularly the case, for cementation and hardening and various chemical actions begin almost from the time of deposition. It is, however, not the custom to refer to these changes as metamorphism.

Many mineral deposits have undergone great changes from their original conditions. They may have been reached by igneous metamorphism and thus a coal bed transformed into anthracite or a bed of limonite into magnetite. Or they may have been sheared or crushed during regional metamorphism Or, most common of all cases, they may have been altered by surface waters. Such oxidizing surface waters, as well as similar waters at somewhat greater depth, when they have parted with their free oxygen, produce peculiar modifications and often most important enrichments.

A CLASSIFICATION OF MINERAL DEPOSITS[1]

I. Deposits produced by mechanical processes of concentration. (Temperature and pressure moderate.)

II. Deposits produced by chemical processes of concentration. (Temperature and pressure vary between wide limits.)

- A. In bodies of surface waters.
    - 1. By interaction of solutions.
        - *a*. Inorganic reactions.
        - *b*. Organic reactions.
    - 2. By evaporation of solvents.

      Temperature, 0° to 70° C. ±
      Pressure, moderate to strong.
- B. In bodies of rocks.
    - 1. By concentration of substances contained in the geological body itself.
        - *a*. Concentration by rock decay and residual weathering near surface.

          Temperature, 0°–100° C. ±
          Pressure, moderate.
        - *b*. Concentration by ground water of deeper circulation.

          Temperature, 0°–100° C. ±
          Pressure, moderate.
        - *c*. Concentration by dynamic and regional metamorphism.

          Temperature up to 400° C. ±
          Pressure, high.
        - *d*. Zeolitization of surface lavas.

          Temperature, 50°–300° C. ±
          Pressure, moderate.
    - 2. Concentration effected by introduction of substances foreign to the rock.
        - *a*. Origin independent of igneous activity.

          By circulating atmospheric waters at moderate or slight depth.

          Temperature, to 100° C. ±
          Pressure, moderate.
        - *b*. Origin dependent upon the eruption of igneous rocks.
            - a. By hot ascending waters of uncertain origin, but charged with igneous emanations.
                - 1. Deposition and concentration at slight depth.

                  Temperature, 50°–150° C. ±
                  Pressure, moderate.
                - 2. Deposition and concentration at intermediate depths.

                  Temperature, 150°–300° C. ±
                  Pressure, high.
                - 3. Deposition and concentration at great depth or at high temperature and pressure.

                  Temperature, 300°–500° C. ±
                  Pressure, very high.

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[1] Presented before the Geological Society of Washington, May 10, 1911.

- b. By direct igneous emanations.
  - 1. From intrusive bodies. Contact metamorphic deposits and allied veins; Temperature, probably 300°–800° C. ± Pressure, very high.
  - 2. From effusive bodies. Sublimates, fumaroles. Temperature, 100°–400° C. Pressure, atmospheric to moderate.

C. In magmas, by processes of differentiation.

- *a.* Magmatic deposits proper. Temperature, 700°–1500° C. ± Pressure, very high.
- *b.* Pegmatites. Temperature, about 575° C. ±. Pressure, very high.

CHAPTER XV

DEPOSITS FORMED BY MECHANICAL PROCESSES OF TRANSPORTATION AND CONCENTRATION; DETRITAL DEPOSITS

INTRODUCTION

Weathering tends to destroy rocks and mineral deposits by disintegration and chemical decomposition. In part, new minerals, like kaolin and limonite, form; in part, the more resistant minerals, like quartz, gold, platinum, magnetite, cassiterite, and garnet, are set free in individual grains. Erosion now steps in and the detritus is swept down the slopes and into the water channels. Mechanical separation in running water or along sea or lake beaches sorts the detritus according to specific gravity and size of grains. The heaviest particles, as those of gold, magnetite, and garnet, tend to collect in the lower part of the assorted detritus; the minute and easily moved scales of clayey substance are carried far but ultimately deposited as sedimentary beds; the colloids are coagulated by the electrolytes in the sea water.

DETRITAL QUARTZ DEPOSITS

The quartz grains are often accumulated as beds of almost pure quartz sands. These are used extensively as ingredients in pottery and glass, also for abrasive purposes in sawing soft rocks, such as marble. Such sands should contain 99 per cent. silica. Somewhat argillaceous quartz sands without carbonates and carrying 80 to 90 per cent. silica are used as molding sands and are mined on a large scale, though occurring in thin beds.[1] When compacted by pressure and by cementation the quartz sands are transformed into siliceous sandstones and quartzites and these are used for millstones, whetstones, and grindstones. Comparatively few localities furnish good material.

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[1] E. F. Burchard, Requirements of sand and limestone for glass-making, *Bull.* 285, U. S. Geol. Survey, 1906, pp. 473–475.

L. Heber Cole, The occurrence and testing of foundry molding sands, *Trans.*, Canadian Min. Inst., vol. 20, 1917, pp. 265–291.

With the development of modern methods of grinding the importance of millstones has greatly decreased. Technical and statistical information on these subjects is contained in *Mineral Resources of the United States,* 1916, part 2, under "Abrasive Materials" and "Sand and Gravel" and for 1916, p. 634, where a list of literature may also be found.

In case of very fine-grained whetstones a doubt may exist whether the material is of detrital origin or formed by chemical agencies. The so-called novaculite of Arkansas, the best whetstone known, is a good example of this. It occurs in the Silurian beds of Garland and Saline counties in that State, and is used for what are known, according to color and quality, as Washita and Arkansas stones. The latter are snow-white and are the harder. The rock is much jointed and only small pieces are obtainable. Branner considers this material a metamorphosed chert, while Griswold[1] believes it to be a fine-grained sediment.

DETRITAL CLAY DEPOSITS[2]

The fine material resulting from the decay of rocks is carried away, suspended in water, and deposited in river beds, lakes, and seas as sedimentary clay. The nature of clays is a much discussed subject. Perhaps the best definition is given by G. P. Merrill, who says[3] that the clays are widely diverse in origin and in mineral and chemical composition but have the common property of plasticity when wet and that of induration when dried. Clays are finely comminuted aggregates of hydrous aluminous silicates, detrital quartz and other mineral fragments, often also, iron hydroxide and calcic and magnesic carbonates. The sedimentary clay is therefore to be regarded rather as a rock than as a mineral and its principal use is for structural purposes. The detailed description of these deposits, therefore, does not fall within the scope of this book.

The larger part of the clays are derived from decomposition

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[1] L. S. Griswold, Whetstones and the novaculites of Arkansas, *Ann. Rept.* Arkansas Geol. Survey, vol. 3, 1890.

[2] For more details in regard to the important clay industry the reader is referred to H. Ries, Clays, New York, 1908. Information as to production, etc., is given in *Mineral Resources of the United States,* published annually by the U. S. Geol. Survey. Further notes regarding residual kaolin deposits may be found on pp. 325–328.

[3] G. P. Merrill, Rocks, rock-weathering, and soils, New York, 1897, p. 135.

and hydration of feldspathic minerals; other silicates, however, contribute their share. It has been supposed that the mineral kaolinite ($H_4Al_2Si_2O_9$) is one of the principal constituents of clay. Probably it is present because the formation of kaolinite from feldspars can easily be traced in decomposing rocks at the surface, but in the clays the mineral is so comminuted that it cannot be readily identified. It is known that colloid hydrous silicates of aluminum exist and there are also a number of more or less indefinite compounds of this kind in nature, such as halloysite, smectite, and pholerite. The sedimentary clays rarely approach kaolinite in composition. Kaolinite should contain 46.5 per cent. SiO_2, 39.5 per cent. Al_2O_3, and 14 per cent. H_2O; but by reason of admixture of quartz and undecomposed silicates, the sedimentary clays usually contain much more silica than the amount indicated.

Clays without carbonates generally contain more magnesium than calcium, and potassium exceeds sodium. Titanium often exceeds one per cent. Much of the titanium and potassium is probably present in colloid state. Traces of copper, nickel, lead, zinc and vanadium are sometimes found.

Regarding residual clays derived from the decomposition of rocks in place see p. 325. The clays formed by the action of sulphuric acid on silicates in the oxidized part of ore deposits are described on p. 480.

FULLER'S EARTH[1]

Fuller's earth is the name given to certain sediments of clay-like material, originally used in England by fullers for cleansing cloth of grease. At present this substance is extensively used for deodorizing, decolorizing and clarifying fats and oils; much of it is employed in the refining of petroleum. Its value thus depends upon its adsorbent qualities.

The material occurs in sedimentary beds of Mesozoic, Cenozoic,

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[1] J. T. Porter, Properties and tests of fuller's earth, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 268–290.

T. W. Vaughan, Fuller's earth of Florida and Georgia, *Bull.* 213, U. S. Geol. Survey, 1903, pp. 392–399.

Mineral Resources, U. S. Geol. Survey. Annual publication. Articles by F. B. Van Horn and J. Middleton.

Charles L. Parsons, Fuller's earth, Bureau of mines, *Bull.* 71, 1913.

E. H. Sellards and H. Gunter, *Second Annual Report*, Florida Geol. Survey, 1908–09, pp. 255–290.

and Quaternary age, but a similar material is also derived from the weathering of basic igneous rocks. Microscopic examination gives little evidence of its origin; in color it ranges from gray to dark green; it possesses little or no plasticity. The chemical analysis also has little value in determining its quality. J. T. Porter believes, and probably justly, that the material owes its quality to the adsorbent power of colloid hydrous aluminium silicates.

The analyses show that the silica varies between 47 and 75 per cent., alumina from 10 to 19 per cent., lime from 1 to 4 per cent., magnesia from 2 to 4 per cent., ferric oxide from 2 to 10 per cent., and combined water from 5 to 21 per cent.

In Gadsden County, Florida and Decatur County, Georgia, it occurs in Tertiary strata and is mined in open pits; in Arkansas it is obtained from weathered basic dikes. The further preparation includes drying, grinding and bolting to sizes from 30 to 100 mesh per inch. Very fine material clogs the filter presses. The domestic production amounts to about 70,000 short tons and about 17,000 tons were imported. Florida yields most of the total domestic production. The price of the Florida material is about $10 per ton.

PLACER DEPOSITS

Origin and Distribution.—The heavier and less abundant minerals in the rocks are the most resistant to decomposition and when the weathered rock is eroded and sorted by water they usually become concentrated in the lower parts of the sand and gravel beds. The gold-bearing gravels, which form an important source of supply of this metal, were called placers[1] by the early Spanish miners of this continent, and this name is probably the best that can be adopted for deposits of this class. Instead of gold the valuable mineral may be cassiterite, magnetite, monazite, diamonds, or other precious stones. Other terms have been employed, as "gravel deposits" or "gold-bearing gravels," or "alluvial deposits"—all equally objectionable, for the material may be sand instead of gravel, and it may be deposited along the ocean beach instead of in watercourses.

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[1] Derivation uncertain: Placer, pleasure; Plaza, place. Stelzner (Die Erzlagerstätten, p. 1261) says placer is a local Spanish term for sand bank. The Germans use "Seife," meaning washings. In French the word "alluvions" is often used.

The processes of erosion and concentration have been active since earliest geologic time, and so we may have detrital deposits or placers of differing ages. Land deposits are, however, usually thin and easily removed and thus placers of pre-Tertiary age are comparatively rare.

In the formation of placers nature simply employs in her own leisurely way the processes of crushing and concentration which we use in ore dressing. The rocks are broken and comminuted by the expansion due to alternating heat and cold; by the growth of plants; or by the impact of sliding and water-carried rocks; or by the grinding action of ice; or finally by chemical decomposition and hydration. The products are concentrated in water courses or along shores by running water or in ocean currents by motion similar to that on tables and jigs. Spherical particles of different substances fall in water at a rate proportional to their weight divided by the resistance. As the resistance is proportional to the area exposed, a fragment of quartz the size of a pea will fall much more slowly than a piece of gold of the same size. It will in fact be carried along easily in a current of water in which a piece of gold of the same size will sink instantly. Thus the specific gravities of the valuable minerals play a prominent part in the formation of placers. The specific gravity of the more important substances is as follows: Quartz, 2.64; feldspar, 2.55 to 2.75; ferromagnesian silicates, 2.9 to 3.4; garnet, 3.14 to 4.13; diamond, 3.54; corundum, 4.0; monazite, 5.0; magnetite, 5.0; cassiterite, 6.4 to 7.1; gold, 15.6 to 19.33; platinum, 14.0 to 19.0 (21 to 22 when chemically pure).

The shape of the particles is also of importance. Flaky minerals, like molybdenite, scaly gold, or specularite, are difficult to concentrate in spite of their high specific gravity.

GOLD PLACERS

Introduction.—Gold is the most important placer mineral. Roughly speaking, about $70,000,000 out of a world's production of about $450,000,000 are derived from Tertiary or Quaternary placer deposits; discoveries in Alaska and the Northwest Territory have lately increased the output. Gold placers as a rule are easily discovered and worked; the supplies of old and long-settled countries were generally long ago exhausted. Bohemia, Italy, Spain, and Hungary, now almost barren of placers, once furnished their share. New deposits are usually discovered on the outskirts of civilization, as in Brazil in the eighteenth century, in Australia and California during the middle of the last century, and in Alaska and Siberia to-day. The production of placer gold in the United States, including Alaska, in 1897 was $7,800,000; in 1916 it was $22,882,000, the increase being due to the recently discovered placers of Alaska and to the development of the dredging fields in California. Practically all this gold comes from Quaternary and Tertiary placers, some dating back as far as the Eocene. A small quantity is obtained from Cretaceous conglomerates in Oregon and northern California. Permian gold-bearing conglomerates occur in Bohemia, according to Posepny.[1] Permo-Carboniferous conglomerates containing detrital gold have been described by Wilkinson from New South Wales.[2] In most cases the gold content of these older conglomerates is small and they can rarely be profitably worked. Probably the best example of ancient placers is furnished by the Cambrian basal conglomerate of the Black Hills, South Dakota, which unconformably covers the pre-Cambrian schists and gold-bearing quartz veins. It was first described by W. B. Devereux[3] and later by J. D. Irving.[4] This conglomerate, which is from 2 to 30 feet thick and is overlain by quartzite, carries in places much gold of unquestionably detrital origin, as indicated by the rounded grains, and has been profitably worked in several mines. The gold was derived from the erosion of auriferous lodes in the pre-Cambrian rocks and was deposited in depressions along the old shore line. In part the gold-bearing conglomerate is cemented by pyrite, which probably also contains some gold. Maclaren[5] believes that the scarcity of economically important deposits of detrital gold in older formations is due to its solution, in depth, by alkaline solutions. There is little evidence in support of this view.

Origin of Placer Gold.—In primary deposits gold is mainly contained in veins, lodes, or shear zones and these appear in

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[1] Genesis of ore deposits, 1902, p. 163.

[2] *Idem*, p. 162.

[3] W. B. Devereux, *Trans.*, Am. Inst. Min. Eng., vol. 10, 1882, pp. 465–475.

[4] J. D. Irving, Economic resources of the northern Black Hills, *Prof. Paper* 26, U. S. Geol. Survey, 1904, pp. 98–111.

[5] J. M. Maclaren, Gold, London, 1908, p. 90.

rocks of many different kinds. It is often stated that gold is distributed as fine particles in schists and massive rocks and that placer gold in certain districts is derived from this source. Most of these statements are not supported by evidence, though it is not denied that gold may in rare instances be distributed in this manner. Even in the Yukon region, concerning which such statements have often been made, the origin of the gold from veins, lodes, and shear zones is beginning to be recognized.[1]

The great majority of gold placers have been derived from the weathering and disintegration of auriferous veins, lodes, shear zones, or more irregular replacement deposits. These primary deposits were not necessarily rich and may not be profitable to work. In many regions the rocks contain abundant joints, seams, or small veins in which the gold has been deposited with quartz.

Eluvial Deposits.—Gold placers may be formed by rapid erosion of hard rocks, but such placers are not often rich and highly concentrated. In the great placer regions the concentration has generally been preceded by an epoch of deep secular decay of the surface. It has been supposed by many that this deep rock decay is peculiar to the tropics, but this is not correct. The process has been active in the southern Appalachian States, in California, and even in Alaska, as well as in countries like the Guianas and Madagascar. When the outcrops of gold-bearing veins are decomposed a gradual concentration of the gold follows, either directly over the primary deposits or on the gentle slopes immediately below. The vein when located on a hillside bends over (Fig. 73) and disintegration breaks up the rocks and the quartz, the latter as a rule yielding much more slowly than the rocks; the less resistant minerals weather into limonite, kaolin, and soluble salts. The volume is greatly reduced, with accompanying gold concentration. The auriferous sulphides yield native gold, hydroxide of iron, and soluble salts. Some solution and redeposition of gold doubtless take place whenever the solutions contain free chlorine. The final result is a loose, ferruginous detritus, easily washed and containing easily recovered gold. This gold consists of grains of rough and irregular form and has a fineness but slightly greater than that of the gold in the primary vein. Stelzner has applied to such residual concentrations, which may be worked like ordinary

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[1] A. H. Brooks, The gold placers of parts of Seward Peninsula, Alaska, *Bull.* 328, U. S. Geol. Survey, 1908, pp. 108 *et seq.*

placers, the term *eluvial* gold deposits. Their occurrence is illustrated in Fig. 73.

In the gold region of the southern Appalachian States the decomposition of the country rock, which generally is a schist, may reach a depth of 100 feet or more.[1] The decomposed material of the auriferous veins slides downhill, mixing with the weathered rock, and during this process the gold in part sinks deeper into the detritus. This has given rise to a peculiar system of mining by which the whole mass is washed by the hydraulic method and the more resistant quartz boulders crushed in a stamp mill with coarse mesh. This has been practiced at Dahlonega and is often called the Dahlonega system. Similar deposits were worked in California, particularly in Eldorado county, and are here called "seam diggings" from the fact that the gold occurs disseminated in quartz seams traversing a certain belt of schists. Such deposits frequently occasion legal contests owing to the uncertainty whether they should be considered as placers or as mineral-bearing veins.

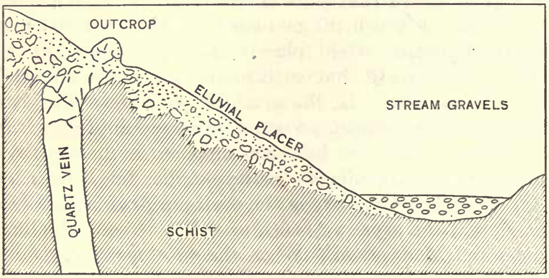

FIG. 73.—Diagram showing development of eluvial and stream placers.

In certain regions of Brazil[2] the schists and gneisses are covered by auriferous detritus accumulated in place. Another example is the "Tapanhoancanga" of the same country. This is a bed of residual or lateritic iron ore up to 10 feet thick covering the

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[1] G. F. Becker, Reconnaissance of the gold fields of the southern Appalachians, *Twenty-sixth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1895.

[2] O. Derby, Peculiar modes of occurrence of gold in Brazil, *Am. Jour. Sci.*, 3d ser., vol. 28, 1884, p. 440.

O. Derby, Notes on Brazilian gold ores, *Trans.*, Am. Inst. Min. Eng., vol. 33, 1892, pp. 282–283.

underlying hematite schist and containing gold throughout. The gold probably occurs in veinlets in the schists and the gold-bearing detrital material has been concentrated from a considerable thickness of schist weathering in place.

Excellent examples of eluvial deposits are reported from Dutch, British, and French Guiana,[1] though ordinary stream placers are the most common deposits in these countries. Over a great part of this gold-bearing territory secular decay of crystalline rocks has resulted in a deep mantle of ferruginous clayey earth—laterite—and in places the gold has been concentrated in this material below outcrops of gold-bearing veins. Many of the stream beds are also worked for placer gold, the detritus usually resting on the clayey surface of the compact laterite.

It is stated that many rocks in the Guianas centain gold and that the placer gold is derived from such material; particularly are the basic rocks, diabases and amphibolites, said to be auriferous. This conclusion should probably be accepted with some reserve. It seems more probable that the gold contained in the greenstones is of secondary origin and that here, as elsewhere, granitic intrusions have caused the formation of a series of gold-bearing veins in the surrounding rocks.

Processes of Concentration.—In most cases the cycle has been carried further and the material is not only decomposed, but eroded, transported, and redeposited. This can be effected by wind, by streams, or by the surf of the sea.

Eolian Deposits.—Deposits concentrated by eolian agencies can, of course, be formed only in dry countries where long subaerial decay has paved the way for the work of the dust storms;

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[1] C. G. Dubois, Geologisch-bergmännische Skizzen aus Surinam, Freiberg i. S., 1901, pp. 112.

C. G. Dubois, Beiträge zur Kenntniss der surinamischen Laterit, etc., *Tsch. Min. u. petr. Mitt.*, 22, 1903.

E. D. de Levat, The gold fields of French Guiana, *Mineral Industry*, vol. 7, 1899.

E. D. de Levat, Guide pratique etc. de l'or en Guyane française, Paris, 1898.

A. Bordeaux, *Trans.*, Am. Inst. Min. Eng., vol. 41, 1910, pp. 567–593.

J. B. Harrison, The geology of the gold fields of British Guiana, London, 1908.

J. B. Harrison, in the Reports of the Instit. of Mines, British Guiana.

E. E. Lungwitz, Die Goldseifen von British Guiana, *Zeitschr. prakt. Geol.*, 1900, pp. 203–218.

from the decomposed and crumbled outcrops of the lodes the winds blow away the lighter sand, leaving a mass of coarser detritus which contains the gold. Such wind-born placers have been noted by H. C. Hoover[1] and T. A. Rickard[2] near the croppings of the West Australian gold veins. No examples of this kind are known from the Cordilleran States of America.

Stream Deposits.—Running water is by far the most important agency in the formation of gold placers. First of all, attention must be directed to the high specific gravity of gold, which explains many of the puzzling features of the placers. Placer gold is six or seven times as heavy as the most common accompanying minerals—feldspar and quartz—and it settles to the bottom in flowing water with surprising rapidity. It is almost impossible to lose a particle of gold, of the value of one cent, in a miner's pan; it sinks immediately to the bottom of the gravel and sand after one or two preliminary shakes in water. Once lodged at the bottom it stays there, in spite of shaking and rotating. This illustrates the fundamental fact that the gold is mainly on the bed-rock. The rapid settling of the gold accounts for the partial failure of some devices for placer mining, particularly the clam-shell and the suction dredges.

The ease with which some concentration, according to the specific gravity, is effected is shown by the well-known fact that in powdered samples of ore, as well as in dumps at the mine, a settling of the heavier ore particles toward the bottom can often be observed.

Suppose we have a gold-bearing quartz vein deeply altered by rock decay; now let the region be raised say 500 feet by one of these slow oscillations which so commonly affect the crust. A river has excavated a valley to the corresponding depth in this elevated plateau, and this valley—under the influence of a pause in the elevating movement—becomes filled with gravels to a width of about 100 feet. Let a tributary gulch with steep grade be cut back into the plateau to the gold deposit (Fig. 74); when the gulch reaches it the eluvial deposit will be carried down by sliding and washing; the clay and limonite are rapidly removed in suspension; the angular gravel of quartz and rock, grinding the

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[1] H. C. Hoover, The superficial alteration of Western Australian ore deposits, *Trans.*, Am. Inst. Min. Eng., vol. 28, 1899, pp. 762–763.

[2] T. A. Rickard, The alluvial deposits of Western Australia, *idem*, pp. 480–537.

fragments of gold between them and on the bed-rock, will be moved downward, the fine grains in suspension, the coarser ones dragging and rolling on the bottom. There is little deposition; the transporting power is great and in flood time the whole gravel mass, of no great depth, will probably be in motion. Heavy gold nuggets may lodge in the lee of little ridges. The gold settles rapidly; most of it, continually hammered and slowly shaping itself in flat, smooth grains, will be dragged down stream and finally reach the edge of the flood-plain in the river.

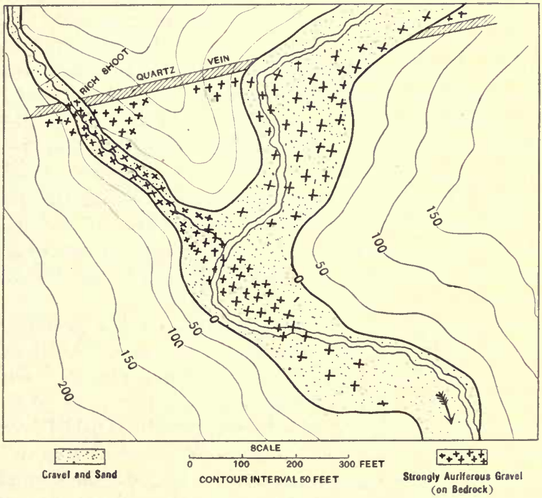

FIG. 74.—Plan of quartz vein and placers below it, illustrating the development of pay streaks.

At this place the larger part of the gold stops. It is not washed out with the sand and gravel but stays on the bed-rock near the margin. The finer particles will, of course, be carried out a little distance, but they soon sink into the water-filled gravel after the manner of grains of heavy ores in concentrating jigs. Just as in the gulch the whole mass of detritus is transported, so it is thought that in larger streams the body of water-soaked gravel and sand works downstream very slowly. During this process the lighter gold contained in the detrital material also works forward and downward, gradually joining the nuggets or coarser pieces, which have already reached their final resting ground.

This mode of operation contains the key to the genesis of the placers. It is not to be expected that the coarse and ordinary fine gold will be carried out into the middle of wide flood-plains. As the flood-plain widens it will cover the accessions of gold along its margin, and the final result will be a streak of rich gold-bearing gravel, resting on the bed-rock and extending downstream deep underneath the surface. When this is traced upstream the primary deposit, the vein, will be found. The actual occurrences of course show infinite variation. Let us assume that, as happens in the Creswick district in Victoria, Australia, a broad stream with moderate grade crosses a deeply decomposed belt of soft slate containing an abundance of small veins or stringers of quartz with native gold, and that in addition a fair balance between transportation and deposition persists for a long time. The result will be a gravel deposit, only a few feet deep, but with an abundance of gold concentrated on the bed-rock over the whole width of the stream. Each freshet is sufficient to churn up and move forward the whole mass of gravel, continually adding to the concentrated gold on the clayey bed-rock.

Again, we may assume extremely active erosion, as is the case in the Sierra Nevada of California. Canyons several thousand feet in depth have been cut in an uplifted plateau, veritable trenches or sluice boxes, the grade of which is from 60 to 150 feet per mile. Stretches of wild gorges with polished bottoms alternate with stretches of less grade where shallow gravel accumulates. These canyons receive for long distances an abundant supply of gold of all sizes from older hill gravels or from decaying quartz veins. The result will be that but little gold will lodge in the gorges, while extremely rich shallow gravel bars will accumulate in the convex stream curves (Fig. 75). Gradient, volume, and load usually vary in the same stream so that deposition may be going on in one part of its valley and erosion in another. Continued corrasion of the stream-bed results in deepening the canyon and leaving the bars as elevated benches. The miners of 1849 first found these bars and worked them. In searching for the source of the gold they soon found a trail of metal leading up the gulches to great masses of older gravels on the hills, 2,000 to 3,000 feet above. These gravels were washed by the hydraulic method; immense masses of tailings with a little gold were carried down to the rivers, totally overloading them. After the prohibition of hydraulic mining the streams gradually resumed active transportation. The whole gravel mass moved slowly downstream and a gradual reconcentration on the bed-rock took place. The tailings deposited became enriched and will ultimately be reworked.[1]

The torrential floods of the canyons scarcely permitted the lodgment of fine gold. This was swept out through the narrow portals into the Sacramento Valley, where the grade of the streams suddenly diminishes. The most minute particles may have been carried as far as San Francisco Bay, but the bulk of the fine gold lodged in the flood-plains within a few miles of the mouth of the canyons. Easily caught upon the clayey *"false bed-rock"* of a volcanic tuff, this gold, the average particles of which are about 0.3 millimeter in diameter, formed meandering pay streaks at the base of a sandy gravel bed from 10 to 60 feet in depth. Such deposits are now worked by dredging.

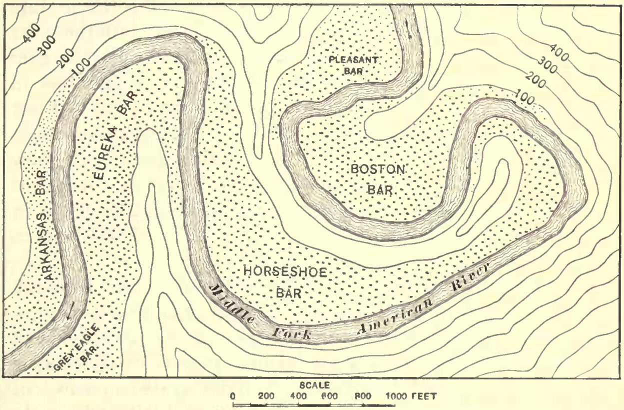

FIG. 75.—Low gravel bars, American River, California, showing placer deposits on inner side of bends. *After R. L. Dunn.*

By an odd paradox, gold is at the same time the easiest and the most difficult mineral to recover. It is divisible to a high

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[1] G. K. Gilbert, Hydraulic-mining débris in the Sierra Nevada, *Prof. Paper* 105, U. S. Geol. Survey, 1917.

degree and owing to its insolubility the finest particles are preserved. A piece of gold worth one cent is without trouble divisible into 2,000 parts, and one of these minute particles can readily be recognized in a pan. In extreme subdivision the gold acquires a scaly, flat form, being known as flour gold or flake gold, is carried away very readily by water, and does not sink easily in sand or gravel. In part the flour gold is suspended by air films, and it can be carried away in rivers of moderate grade for hundreds of miles. The gold occurring in the sand bars of Snake River, Idaho, is a good example of this.[1] It will settle in thin pay streaks at bars and other favorable places, but the next freshet will probably destroy the sand bars and sweep the gold away. This accounts also for the distribution of fine gold in great masses of gravel beds—for example, in the wash 600 feet thick deposited by glacial streams at Tacoma and other places on Puget Sound. Almost every pan of this gravel will show a "color," but the material contains only a fraction of a cent per cubic yard. The fine colors along the Columbia River in northeastern Washington range in value from less than 0.0005 to 0.02 cent, the average being about 0.002 cent.[2]

The much-discussed concentration of gold on the bed-rock seems, then, to be due partly to the natural jig-like movement in moderately deep gravels,[3] during long-continued conditions of fair balance between loading and erosive power; partly to slow forward and downward motion of heavier gravel masses,[4] of which exact measurement as yet is lacking, and last and largely to the fact that heavier gold will not be carried out into the gravel flats of rivers of gentle grade—the only ones that have extensive flood-plains—but is immediately deposited on the marginal bed-rock of the gradually deepening and widening gravel plains.

The best conditions for the concentration of gold are found in moderately hilly countries where deep secular decay of rocks has been followed by slight uplifts. Subsequent slight elevations would easily produce re-sorting and enrichment of the gravels. In regions of gold placers the richest material is usually pro-

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[1] J. M. Hill, Gold of the Snake River, *Bull.* 620, U. S. Geol. Survey, 1916, pp. 271–294.

[2] A. J. Collier, *Bull.* 315, U. S. Geol. Survey, 1907, p. 61.

[3] F. Posepny, Genesis of ore deposits, New York, 1902, p. 154.

[4] T. A. Rickard, *Min. and Sci. Press*, Aug. 15, 1908.

duced by repeated reworking of gold-bearing gravels by nature. Each reworking increases the richness of the gravels, eliminates easily decomposed pebbles, and finally results in a gravel of the hardest, most resistant rock—quartzite or quartz. Quartz is the common gangue mineral in gold regions; hence the prevalence of "white gravels" or "white channels," almost exclusively composed of white quartz pebbles.

CLASSIFICATION OF FLUVIATILE AND MARINE PLACERS

According to their occurrence the placers may be conveniently divided as follows.[1]

PLACERS CLASSIFIED

Present topographic cycle	Past cycles, elevated	Past cycles, depressed
1. Gulch and creek gravels.	1. High creek gravels.	1. Deep creek gravels.
2. River and bar gravels.	2. { Bench gravels. Hill gravels or high river gravels.	2. Deep river gravels.
3. Gravel plains.	3. Elevated gravel plains.	3. Depressed gravel plains.
4. Beaches.	4. Elevated beaches.	4. Depressed beaches.

Examples of present gulch, creek, and river gravels are not difficult to find; they occur in all gold-bearing regions where erosion is active and where precipitation is abundant enough to cause the sorting and carrying forward of the gravels in the stream beds. In the upper parts of the stream courses the gravel will be coarse and semiangular; in the lower parts the sands increase and the pebbles are smoother. Where the rivers emerge from their narrow valleys and spread with gentle grade over flood-plains, more extensive sand and gravel beds will accumulate, generally, however, with less gold than in the more confined part of the course. Some of the fine gold may reach the sea and is concentrated by the surf and the oblique shore currents into thin pay streaks on the sandy beach.

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[1] See also A. H. Brooks, The gold placers of parts of Seward Peninsula, Alaska, *Bull.* 328, U. S. Geol. Survey, 1908, p. 115.

Marine Placers.—Beach placers occur along many shores and are often produced by concentration from a sea bluff or elevated gravel plain. The beach at Nome, Alaska (Fig. 76), is a narrow strip about 200 feet wide, from which over $2,000,000 in fine gold has been washed; the flaky gold averaged 70 or 80 colors to the cent.[1] Two older elevated beach lines are found farther inland. The beach gold of the Oregon and California coasts is much finer, the colors ranging from 100 to 600 to the cent.

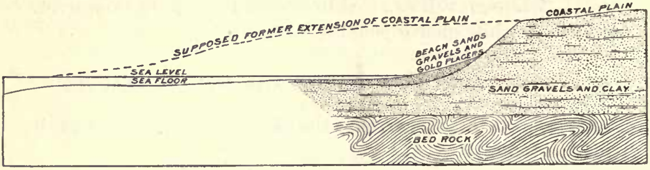

Fig. 76.—Diagrammatic section illustrating development of beach placer. *After A. J. Collier and F. L. Hess, U. S. Geol. Survey.*

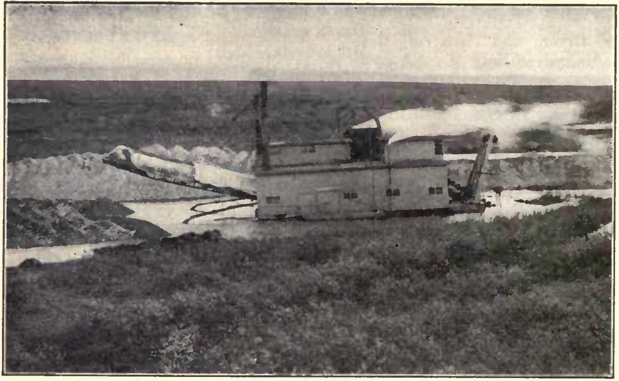

Fig. 77.—Gold dredging on the Solomon River, Alaska. *After P. S. Smith, U. S. Geol. Survey.*

Buried Placers.—Subsidence or overloading may cause the placers to be deeply covered by barren detritus. Many of the streams of Alaska, particularly in their lower reaches, are thus covered; the process of concentration is stopped, the present

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[1] A. J. Collier and F. L. Hess, *Bull.* 328, U. S. Geol. Survey, 1908, pp. 140–228.

watercourses having insufficient grade to effect the transportation of detritus. Fig. 77 shows the dredging operations on the Solomon River, Alaska. The depth of the gravel in the river bottom is about 20 feet. Fig. 78 shows a diagrammatic section of the Oroville dredging ground, Butte County, California. The depth of the gravel is about 30 feet. At Fairbanks, Alaska, according to Prindle,[1] the placers occur in tributaries of moderate length, which flow in open valleys; some of the deposits are as much as 300 feet deep. The pay gravels, in part subangular, lie on the bed-rock and are from a few inches to 12 feet in thickness; these are covered by 10 to 60 feet of angular wash, evidently accumulated rapidly without opportunity for concentration, and above this rests a thick deposit of muck over which the sluggish streams pursue their way. The richest gravel worked in 1905, containing from $5 to $10 per cubic yard, occupies pay streaks on the bed-rock 150 to 200 feet wide, considerably less than the average width of the valley bottom. All the gravel on the bed-rock is, however, more or less auriferous. The gold is moderately coarse. Near the head of the stream deposition closely follows cutting and there the deeply buried, more or less permanently frozen pay streaks of the lower valleys merge into the deposits of the present stream activity.

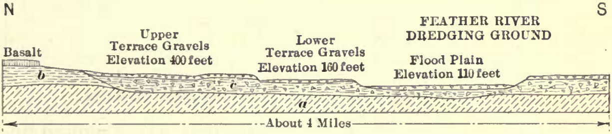

FIG. 78.—Diagrammatic section across Feather River below Oroville, California. *a*, Bed-rock; *b*, Ione formation; *c*, tuffs of Oroville.

On a large scale similar conditions prevailed in Victoria, Australia.[2] Here there existed in Pliocene time an extensive river system with shallow, well washed, and locally extremely rich gravels which were formed during a prolonged time of nice balance between erosion and deposition. The region then

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[1] L. M. Prindle, The Fairbanks and Rampart quadrangles, *Bull.* 337, U. S. Geol. Survey, 1908.

[2] W. Lindgren, *Min. Mag.*, 2, 1905, p. 33.
W. Lindgren, *Eng. and Min. Jour.*, Feb. 16, 1905.
H. L. Wilkinson, *Trans.*, Inst. Min. and Met., 1907, p. 9.
Stanley Hunter, *Mem.* 7, Geol. Survey Victoria, 1909.

became depressed and covered by thick beds of sand and clay. Above this were poured out basalt flows, in places several hundred feet thick (Figs. 79 and 80). The broad valleys remain on the whole as before, but the present streams are weak and have little power of transportation and concentration. The discoveries of gold were made near the sources of the old rivers, where their gravels are near the surface; they were followed upward into the gullies of the slate hills, and downward below the level of the basalt flows. Such were the conditions, for instance, at Ballarat. South of Ballarat certain of the Pliocene stream gravels merge into coastal gravel plains, soon becoming marine in character. Such coastal gravel beds are opened in the Pitsfield mines, where the pay streaks of fine gold, resting on an almost level bed-rock, are worked beneath several hundred feet of sands and gravels.

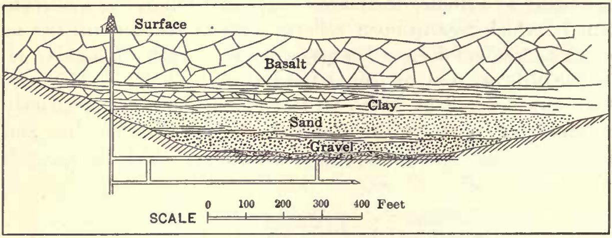

Fig. 79.—Diagram illustrating buried gravel channels (deep leads) of Victoria, Australia, and method of mining these deposits.

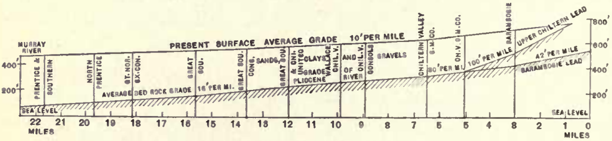

Fig. 80.—Longitudinal section of the Chiltern Valley and Rutherglen deep leads, Victoria, Australia, showing steeper grade of Tertiary river beds.

The Sierra Nevada of California,[1] on the other hand, offers an excellent instance of the result of elevation on gravel deposits. In the early Tertiary the surface of this range was comparatively gentle, and during long periods of rock decay and well-balanced

[1] W. Lindgren, The Tertiary gravels of the Sierra Nevada, *Prof. Paper* 73, U. S. Geol. Survey, 1911.

conditions gold from the quartz veins had become strongly concentrated on the bed-rock of the streams. The deeper gravels were then covered by a considerable thickness of more rapidly accumulated and poorer, but well-washed material, and this in turn by heavy masses of rhyolitic tuffs and andesite breccias so that the old channels were sealed in places by as much as 1,500 feet of superincumbent barren material. The range was elevated by mountain-building disturbances; new rivers were laid out and rapidly eroded cañons to a depth of 2,000 or 3,000 feet. Eventually the old gravels were exposed and now rest as more or less connected remnants on the summits of the ridges between the modern cañons; the heavy gravel masses are worked by the hydraulic method, or the pay streak on the bed-rock is extracted by tunneling operations in the "drift mines" (Figs. 81, 82, 83).

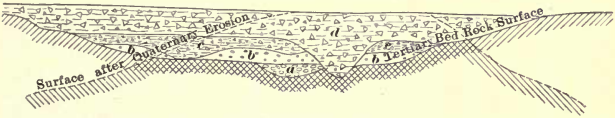

Fig. 81.—Schematic representation of the four principal epochs of Tertiary gravels in the Sierra Nevada. *a*, Deep gravels (Eocene); *b*, bench gravels (Miocene); *c*, rhyolitic tuffs and inter-rhyolitic channel; *d*, andesitic tuffs and intervolcanic channel.

The gold from the destroyed portions of the old channels, together with more set free from the quartz veins during the erosion, accumulated in the modern cañons. Along their slopes benches remain in places, indicating transient accumulations of gravel during the process of cañon cutting.

Somewhat similar conditions exist in some parts of Alaska. Near Nome on the ridges surrounding Anvil Creek are "high gravels" 600 to 700 feet above the present rivers. These gravels, some of which are rich, are the remnants of an old, now almost wholly eroded system of drainage.

In the Klondike also high gravels occur a few hundred feet above the present creeks, the most conspicuous instance being the "White channel," described by McConnell[1] (Fig. 84).

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[1] R. G. McConnell, Klondike gold fields, *Ann. Rept.*, Geol. Survey Canada, 14, 1901.

R. G. McConnell, Report on gold values in the Klondike high-level gravels, *Ann. Rept.*, Geol. Survey Canada, 1907.

Elevated beaches have been mined, for instance, at Nome, where there are two old beach lines 37 and 70 feet above the present level of the ocean. In Santa Cruz County, California, a similar elevated beach was mined for some time. Gold-bearing beach sand occurs all along the Pacific coast from San Diego to Alaska, and in many other parts of the world.

Size and Mineral Association of Placer Gold.—Gold occurs in placers in all sizes, from masses weighing 200 pounds to the most minute flakes. Large nuggets are recorded from California; still larger specimens, weighing as much as 2,184 ounces, were obtained in Victoria, Australia. It is often stated that heavier masses occur in placers than in quartz veins. This is decidedly erroneous. A mass of native gold found in the Monumental mine of Sierra County, California, weighed 1,146 troy ounces, and a quartz vein at Hill End, New South Wales, yielded a specimen which contained about 3,000 ounces. Every one who has had much experience in gold mining has noted the occurrence of thick sheets and masses of gold in deposits of certain kinds—for instance, in the pockety quartz veins of Alleghany, California. Almost all the so-called placer nuggets of unusual size have been obtained from superficial deposits at or just below the croppings of rich veins. This applies to the Ballarat nuggets, weighing from 80 to 160 pounds, which occurred in small steep gulches underneath the basalt flows, but immediately below the extremely rich outcrops of the quartz veins. It also applies to the nuggets of Carson Hill, California, the Poseidon nugget of Victoria (found in 1906 and weighing 953 troy ounces), and other occurrences. Some very rich placer deposits—for instance, those of the Klondike, Yukon Territory, and the Berry mines in

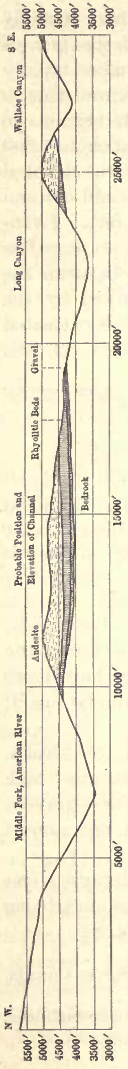

FIG. 82.—Section across the Tertiary Yuba River on the Long Canyon divide, Placer County, California.

Victoria, Australia—contain no specially large pieces of gold. The heaviest nugget found in the Klondike is said to have weighed 85 ounces.

The angularity of the gold is proportional to the distance traveled; the final product is usually a flat, rounded grain from a fraction up to 1 millimeter in diameter. Occasionally crystallized gold is found in placers, but this is unusual and indicates close proximity of the primary deposit.

There is probably no authenticated case of crystallized gold occurring in the gravels of larger water courses where there has been long transportation, and this is assuredly a strong argument against the assumption that such crystals are formed by secondary processes in the gravels.

Fragments of quartz often adhere to the gold or form part of the rounded nugget. While the quartz pebbles so abundantly found in gold-bearing gravels do not ordinarily contain visible gold, there are many instances of such occurrences—for example, at Elk City and Idaho City, Idaho, and at Dutch Flat and Nevada City, California. Some placer gold, more frequently the scaly variety, is covered by a thin film of silica, manganese dioxide, or limonite, and does not amalgamate easily.

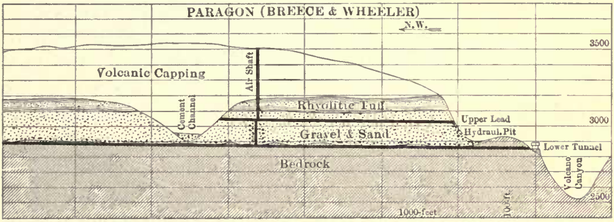

FIG. 83.—Longitudinal section of "Blue gravel channel," at Breece & Wheeler mine, Forest Hill divide, Placer County, California. *After R. E. Browne.*

The most abundant minerals associated with the gold in placers are magnetite and ilmenite ("black sand"), garnet, zircon ("white sand"), and monazite ("yellow sand"), as well as many others of the heavy minerals occurring in the rocks which contain the primary gold deposits. Cassiterite is common in placers, and some deep gold placers in Victoria contain enough to make it a valuable by-product. Gray platinum and silvery foils of iridosmine are present in small quantities in many California placers adjacent to areas of serpentine. None of the minerals mentioned are ordinarily derived from the gold-bearing veins, but from the surrounding rocks.

Pyrite or marcasite may form in the gravels; sometimes this pyrite contains a little gold, but contamination of the assay samples by the placer gold itself is always a possibility. Again the pyrite may be clastic and derived from the surrounding rocks, for pyrite does not seem to oxidize readily in running water; or, as near Nevada City, in the Harmony channel, the gravel may contain undecomposed pyrite, rich in gold, and derived directly from the primary veins over which the water course flowed.

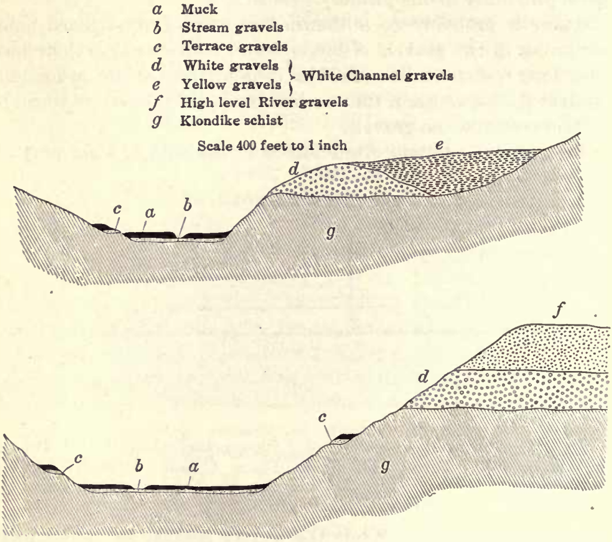

FIG. 84.—Sections across Bonanza Valley, Yukon Territory, showing several types of gravel deposits. *After R. G. McConnell, Geol. Survey Canada.*

Other occasional associates of gold, probably derived from its primary deposits, are silver in nuggets (Alaska), native bismuth

(Queensland and Alaska), native amalgam, palladium-gold, native copper, and cinnabar. The presence of native lead has usually been explained by accidental admixture of hunter's shot, but J. Park[1] asserts that there is an instance of its undoubted presence in gravel, the lead containing a skeleton of native gold.

Fineness and Relation to Vein Gold.—The fineness of placer gold (or parts of gold per thousand) varies from about 500 up to 999. Silver is always alloyed with the gold, but other metals are rarely prominent; copper is occasionally present. While vein gold may have a fineness of 997 to 999, this is exceptional; far more commonly its fineness ranges from 500, which corresponds to electrum, to about 800 or 850. This is assumed to be determined on specimens of native gold, for it is obviously not fair to take the usually lower figure of the retorted bars, which become admixed with impurities during amalgamation. The placer gold in any district will usually be of higher grade than the vein gold, and its fineness increases with the distance transported and with the decreasing size of the grains. Thus, while in California the vein gold averages 850 fine, the transported placer gold in the Tertiary channels averages 930 to 950. It has been shown that this increase in fineness is due to the solution of the silver in the alloy in the outer layer of the grains by the action of surface waters. McConnell has proved that in the nuggets from the Klondike the outside actually has a greater fineness than the inside. The loss of silver in the outer part was from 5 to 7 per cent. This interesting result well illustrates the relative insolubility of gold.[2]

Gold in Relation to Bed-Rock.—While the bulk of the gold usually rests on the bed-rock or within a foot or two of it, this is not an invariable rule. In some gravels the coarser gold is occasionally scattered through the lower 4 to 20 feet. But it is never, except in minute quantities, distributed equally through a

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[1] J. Park, Mining geology, London, 1907, p. 18.

[2] W. B. Devereux, The occurrence of gold in the Potsdam formation, *Trans.*, Am. Inst. Min. Eng., 1882, pp. 465–475.

Ross E. Browne, Colorado placer gold, *Eng. and Min. Jour.*, vol. 59, 1895, pp. 101–102.

W. Lindgren, The gold belt of the Blue Mountains of Oregon, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, p. 637. See also *Prof. Paper* 73, U. S. Geol. Survey, 1911.

R. G. McConnell, Report on gold values in the Klondike high-level gravels, Geol. Survey Canada, 1907, p. 979.

great thickness of gravels. An excellent instance is McConnell's section of the White channel deposit in the Klondike. (See Fig. 84.) The washed gravel is here 150 feet thick. The gold content of the gravel is as follows:

0– 6 feet above bed-rock, $4.13 per cubic yard.
6–12 feet above bed-rock, $0.18 per cubic yard.
12–18 feet above bed-rock, $0.047 per cubic yard.
18–24 feet above bed-rock, $0.04 per cubic yard.
24–30 feet above bed-rock, $0.034 per cubic yard.
30–36 feet above bed-rock, $0.032 per cubic yard.
36–42 feet above bed-rock, $0.032 per cubic yard.
42–48 feet above bed-rock, $0.045 per cubic yard.
48–54 feet above bed-rock, $0.025 per cubic yard.

From 54 feet above bed-rock the quantity of gold contained per cubic yard gradually and steadily diminished to $0.006 at the top. There is only fine gold in the upper gravels. A local enrichment has taken place on false bed-rock, a clayey stratum above the real bed-rock; here the gold is much coarser than directly above or below, but finer than on the bed-rock. Occasionally rich gravel may be found a few feet above bed-rock while it is less rich immediately on it.

Coarse and moderately coarse gold moves very slowly. McConnell found, for instance, that the White channel, where intersected by gulches, has left almost the whole amount of its gold in these immediately below the place where the trenching has occurred. In some cases the horizontal movement scarcely equaled the vertical.

Smooth, hard bed-rock is poorly adapted to retain the gold; when it is somewhat clayey and decomposed much better results are obtained. Schists and slates make good bed-rock when decomposed, especially when they strike parallel to the channel. Serpentine forms a smooth and unsatisfactory bed-rock.

Gold works down into bed-rock in a most surprising way. In hard rock it settles into the most minute crevices. In soft rock it burrows to a depth of 1 to 5 feet, so that it is always necessary to mine this amount of the bed-rock. In limestone, irregular solution cavities contain the detrital gold, and these sometimes descend to a depth of 50 feet or more. Compact clay is good bed-rock, also clayey sandstone and clayey volcanic tuffs, the occurrence of the latter being exemplified in the Oroville dredging grounds, in California.

In glacial till and moraines there has been little opportunity for concentration, and unless the primary vein deposits were unusually rich, these gravels are of little value; the gold contained in them may, of course, be concentrated by glacial streams working over the morainal detritus.

Grade of Auriferous Watercourses.—All kinds of grade occur in watercourses containing gold-bearing gravels. In steep creeks the grade may be many hundred feet per mile, but the placers in these are usually poor. California rivers, in the Sierra Nevada, have grades of 50 to 100 feet or more per mile. Many of these have been extremely rich where gravel bars have had an opportunity to accumulate. The White channel, in the Klondike, has a grade of about 30 feet per mile. Many of the present Alaskan streams have a grade of 100 to 150 feet per mile. In the principal Tertiary channels of Victoria, Australia, low grades down to 20 feet prevailed.

In depressed or elevated channels of past epochs, as in California, Victoria, and the Klondike, changes of original grade must be considered. This is best established in the California channels, which now have grades of 100 to 150 feet, whereas the original streams had much less, the increase being due to the westward tilting of the Sierra Nevada. The best results of gold concentration are probably obtained in rivers of moderate grades, perhaps 30 feet per mile, under nicely balanced conditions of corrasion and deposition. Whenever overloading and active deposition take place concentration of coarse gold ceases. On the other hand, where erosion is rapid conditions for rich placers are less favorable, unless, as in the present streams of the Sierra Nevada, the gold supply is unusually abundant.

The Pay Streak or "Run of Gold."[1]—Except in smaller creeks the distribution of the gold in a gravel bed is far from regular. There is usually gold on the bed-rock over the whole area of the stream bed, but the richer part makes a narrower streak which follows a devious course, distinctly affected by the character of the bed-rock, sometimes splitting and re-forming, following first one side, then crossing diagonally to the other side. It is not necessarily in the deepest depression or gutter. Fig. 85 shows this devious course of the pay streak in comparatively shallow gravels at Maryboro, Victoria. It is clearly independent

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[1] J. B. Tyrrell, The law of the pay streak in placer deposits, *Trans.* Inst. of Min. and Met., London, May 16, 1912, *Min. and Sci. Press*, June 1, 1912.

of the course of the present small stream. In broader gravel plains, of which the Homebush and Pitsfield Tertiary placers of Victoria are examples, the "run of gold" follows a distinct and well-defined course on an almost level country rock. All this shows clearly enough the impossibility of the view that the gold was first uniformly distributed through the gravels and then gradually settled to the bottom under the influence of gravity.

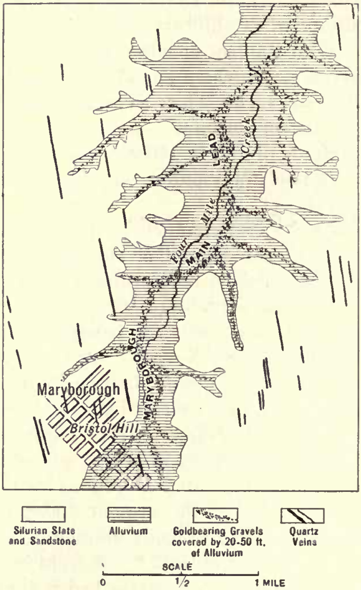

FIG. 85.—Map showing position of pay streak in alluvial gravels of Maryboro, Victoria. *After S. B. Hunter.*

These pay streaks assuredly indicate epochs of well-balanced and long-maintained conditions during which the gravels could accumulate to only moderate depths and were at all times water-soaked and in a condition of slow movement. With more abundant loading of detrital material the gold-transporting power of the stream diminishes at a very rapid rate.

Solution and Precipitation of Gold.—Many of the earlier observers, such as Genth, Lieber, Selwyn, Laur, Egleston, C. Newbery, and Daintree, concluded from observations in various parts of the world that placer gold, and particularly the large nuggets, has been deposited by circulating solutions. At present the mechanical derivation of the gold seems established beyond all doubt, although under exceptional circumstances some solution and redeposition may have taken place.[1] Even now, however, some writers, like J. M. Maclaren,[2] are inclined to place much emphasis on this secondary deposition. It is probable, nevertheless, that this process is absolutely insignificant from an economic point of view. Nuggets, when cut and polished, almost always show a granular structure perfectly in accordance with vein gold. Liversidge, in a long series of experiments, found only two specimens (both from New Guinea) which showed a concentric structure indicative of concretionary deposition. Very rare instances are quoted of quartz pebbles with dendritic films of gold[3] or of nuggets with minute gold crystals on their surface.[4] The collection of J. Edman, of San Francisco, contained a small crystal of magnetite coated with a thin film of gold. This came from the Tertiary deposits at Providence Hill, Plumas County, California, and Mr. Edman stated that he had never seen similar occurrences in the modern gravels. It seems to be well established that pyrite reduced by organic material in the gravels may contain some gold and also that the metal is occasionally found at the roots of trees or in the grass roots.

The gold crystallized in minute octahedrons in the clay of Kanowna, Western Australia, is, as Maitland[5] pointed out, immediately above or adjacent to the decomposed croppings of the veins and the occurrence can scarcely be called a placer. The gold which works down into the soft bed-rock of the placers is in all cases, where I have observed it, of clearly detrital origin.

It is stated that the ashes of trees in the gold-bearing region of the Guianas contain an appreciable quantity of gold. Originally asserted by Lungwitz, this has been denied by Dubois and Kollbeck and then reasserted by Harrison on the basis of

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[1] Regarding the older literature, see the text-books of Stelzner, Bergeat, and Beck. In more detail, see Liversidge, *Jour.*, Roy. Soc. N. S. W., vol. 27, 1893, p. 343; vol. 31, 1897, p. 79; vol. 40, 1906, p. 161.

[2] J. M. Maclaren, Gold, London, 1908, pp. 80–86.

[3] R. G. McConnell, *Ann. Rept.*, Geol. Survey Canada, vol. 14, 1901, p. 64–B.

[4] Gordon, *Trans.*, Am. Inst. Min. Eng., vol. 25, 1895, p. 294.

[5] J. M. Maclaren, *op. cit.*, p. 83.

careful investigations.[1] His statement must be accepted, although it certainly taxes the imagination to believe that gold-bearing solutions can exist in a soil together with organic matter. From widely separate parts of the world gold has been reported in the ash of coal, but in this case it may be detrital and contained in admixed sand and clay.

Gold is easily brought into the colloid state and as such it may be transported in solutions of colloid silica. It is very readily precipitated by electrolytes but this mode of solution may account for some cases of secondary gold in placers. According to the most recent investigations[2] gold is soluble in superficial waters only when free chlorine becomes liberated by the interaction of sulphuric acid, sodium chloride, and manganese dioxide, a combination that must sometimes occur in ore deposits subject to oxidation; in the presence of oxidizing pyrite some gold may therefore be taken into solution, as chloride but it would probably not remain long before encountering reducing substances. While gold is slightly soluble in sodium carbonate, sodium sulphide, and other similar compounds, these would not ordinarily be encountered in the waters of the zone of oxidation.

Relation to Primary Deposits.—That placer gold is directly derived by mechanical processes from vein deposits or analogous occurrences is absolutely certain, and examples of convincing character are present everywhere. This does not imply that the primary deposit can be worked at a profit. In most cases the gold is traceable up to the deposit. On this principle the pocket hunter proceeds, panning the detritus and working up hill until the source of the scattered gold has been found. The area in which the detritus occurs has the shape of a triangle, the apex of which is the pocket.

It is a common experience that rivers or creeks crossing a vein or a mineral belt are enriched immediately below it, the coarseness of the gold increasing upstream to the place where the outcrops are crossed. As examples may serve the great accumulations of placer gold in the Neocene gravels of Eldorado County, California, where the Mother Lode crosses them, and the rich channels in upper Nevada County, just below the belt of

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[1] J. B. Harrison, Geology of the gold fields of British Guiana, London, 1908, p. 209.

[2] W. H. Emmons, Enrichment of ore deposits, *Bull.* 625, U. S. Geol. Survey, 1917, p. 305.

quartz veins at Washington and Graniteville. There are fine examples in Victoria, where the gravels are rich only where they cross or follow systems of veins or "reef lines." The White channel of the Forest Hill divide, California, follows a belt of quartz stringers in clay slate. The Idaho Basin[1] presents an excellent instance of large gravel bodies the gold content of which is traceable up to certain auriferous vein systems.

Economic Notes.—The world's annual production of placer gold is about $70,000,000. To this the Alaska and Yukon districts contribute $20,000,000, California $9,000,000, Victoria $2,000,000, and Siberia $18,000,000. While placers are found in almost all gold- and silver-producing regions, Brazil, the Ural Mountains, Siberia, California, Alaska, and Victoria have had by far the greatest total production.

Gold-bearing gravel is often measured by the ton, but more commonly by the cubic yard. Still another measure is by surface area, sometimes by the square foot, in Australia commonly by the square fathom; this is especially applicable to deep mining when only the richest bottom layer is mined; at least 2 feet of gravel and 1 foot of soft bed-rock are extracted, making one square fathom equivalent to a minimum of 4 cubic yards. .

In river bars gravels are worked by wing dams and pits kept dry by simple pumping devices. On a large scale they may be ground-sluiced or washed by the hydraulic method, with the aid of elevators when the natural fall is insufficient.

The elevated gravels of earlier periods are worked in California by tunnels and drifting operations on the bed-rock. The minimum cost of working under the most favorable conditions is 50 cents per cubic yard, but is commonly $1 to $2 per cubic yard; most of the gravels actually worked contain at least $1.50, and often much more. The whole gravel body may be washed by the hydraulic method, when the expense may be reduced to 2 to 5 cents per cubic yard; of course the cost of preliminary work like ditches, etc., is often great. Some creek gravels in the Seward Peninsula contain from $2 to $6 per cubic yard; the width of the deposit may be about 50 feet, the depth 3 to 6 feet.

The depressed gravels of earlier periods are worked by drifting from shafts, as in Victoria, where, however, the preliminary pumping, to permit access, is an extremely heavy expense, often

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[1] W. Lindgren, The mining districts of Idaho Basin and the Boise Ridge, *Eighteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1898, pp. 617–744.

indeed prohibitory. Some of these Australian channels have been extremely rich, the workable portions ranging from $2 to $15 per cubic yard. Some of the channels are in places several hundred feet in width.

Of late, gravels have been extensively worked in California, Alaska, the Klondike, and elsewhere by the dredging process. In California, where this method has reached its highest development, $7,769,000 was obtained from 58 dredges in 1916 from the flood-plains of the rivers at the foot of the Sierra Nevada, and the cost has been reduced from about 10 cents to 3 or 4 cents per cubic yard handled. In Alaska the cost is of course much higher and gravels containing less than 50 cents per cubic yard are rarely worked. The dredge will probably prove to be the most efficient placer-mining machine of the future, replacing the hydraulic method, which offers difficulties in the disposition of the tailings. In 1917 Alaska yielded $2,500,000 from dredging.

Certain gravels in the dry regions of Arizona and northern Mexico are treated by pneumatic concentration in so-called dry washers, but the output of these placers is insignificant.[1]

Yields of placer deposits are often calculated in dollars per lineal foot of channel. Good channels for drifting may produce from $70 to $500 per foot. The richest drift mine worked was probably "Madame Berry" in Victoria, with average width of 450 feet, yielding $1,293 per foot along channel. The two claims below this produced, respectively, $843 and $443 per foot, the last-named channel being mined 1,000 feet wide. The White Channel in the Klondike gave $380 per foot; the Red Point channel in Placer County, California, $72, the width being 120 feet; the American Hill hydraulic mine, Nevada County, 1,000 feet wide, $414; the Nome creeks, Alaska, 50 feet wide, about $100. By drifting operations alone, only a part of the gold will be extracted, say one-fifth to one-half, dependent upon the thickness of overlying gravels.

THE GOLD-BEARING CONGLOMERATES OF SOUTH AFRICA

Of the extensive literature the following principal papers are quoted:

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[1] T. A. Rickard, *Trans.*, Am. Inst. Min. Eng., vol. 28, 1899, p. 480.
F. J. H. Merrill, *Min. and Sci. Press*, July 13, 1912.

G. A. F. Molengraaff, Die Reihenfolge der geol. Form. in Süd Afrika, *Neues Jahrbuch,* 1900, B. 1, pp. 113–119.

F. H. Hatch and G. S. Corstorphine, Petrography of the Witwatersrand conglomerates, etc., *Proc.*, Geol. Soc. S. A., vol. 7, pt. 3, 1904, pp. 140–145.

F. H. Hatch and G. S. Corstorphine, The geology of South Africa, London. 1905.

J. W. Gregory, The origin of the gold in the Rand Banket, *Trans.*, Inst. Min. and Met., London, vol. 17, October, 1907. Also *Econ. Geol.*, vol. 4, 1909, pp. 118–129.

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R. B. Young, The Rand Banket, London, 1917, pp. 125.

See also description by R. Beck, Erzlagerstätten, 2, 1909, pp. 183–200.

F. H. Hatch, The conglomerates of the Witwatersrand in types of ore deposits, San Francisco, 1911, pp. 202–219.

C. B. Horwood, The Rand Banket, *Min. and Sci. Press*, Oct. to Dec., 1913.

E. T. Mellor, The upper Witwatersrand System; the East Rand, *Trans.*, Geol. Soc. S. A., vol. 18, 1915, pp. 11–71.

E. T. Mellor, The conglomerates of the Witwatersrand with discussion, *Trans.*, Inst. Min. and Met., London, vol. 25, 1916, pp. 226–348.

Hugh F. Marriott, Mining on the Rand, *Trans.* Inst. Min. and Met., London, 1918, *Min. and Sci. Press*, July 20, 1918.

The development of the gold-bearing conglomerates of the Witwatersrand district, in the Transvaal, is one of the most wonderful chapters in the history of mining. From an inconspicuous beginning in 1887, the production of these unique deposits has steadily increased. In 1917 the ore production amounted to about 28,000,000 tons, with a yield of $180,000,000. The total production to the end of 1917 exceeds $2,500,000,000, which is more than the total gold production of California, Colorado and Alaska. The average content of the ore has decreased, probably mostly on account of reduction in mining and metallurgical costs, from $12 to $6 or $7 per ton, and, according to Hatch, it is probable that in the future ore of $5 grade will be utilized. The increase in production continued to 1916 but it is probable that the flood tide of output has been reached. A depth of over 5,000 feet has now been reached, and,[1] owing to a favorable geothermic gradient (p. 84), it will be possible to go considerably deeper. The ore is reduced by a combination of amalgamation and the cyanide process; stamp mills and tube mills are the grinding machinery most commonly employed.

South Africa is in the main a plateau of thick sedimentary beds which are poor in fossils and in part of sub-aërial origin.

[1] In 1918 the Village Deep had attained a vertical depth of 5,350 feet and a depth along the dip of 9,800 feet.

The oldest rocks known are the Swaziland crystalline schists and the granites intruded in them. On their eroded surface rest the upper and lower Witwatersrand system of slates, quartzites, and conglomerates, aggregating 19,000 feet in thickness, and on top of these in turn a thick series of volcanic flows, called the Ventersdorp system (Fig. 86).

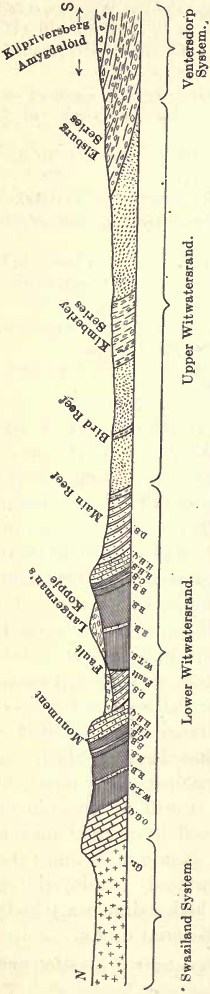

FIG. 86.—Section across the Witwatersrand beds. *After F. H. Hatch.* Length of section about 8 miles.

The age of the Witwatersrand system is not definitely known; it is probably Cambrian or pre-Cambrian. Next higher in the succession of rocks is the Potchefstrom system, including the Black Reef (oldest), Dolomite, and Pretoria series. This again is covered by the Devonian Waterberg system (Table Mountain sandstone of the Cape) and the most recent Karroo system, which is coal-bearing and considered to be of Permo-Carboniferous age Each system is separated by an unconformity from the next.

The Witwatersrand system is folded in a syncline extending about 120 miles east to west and 45 miles north to south. At Johannesburg, at the north side of the syncline, the dip is to the south, steep near the surface, but flattening in depth to about 30°. Faulting is common and there are a number of intrusive diabase dikes, thought to belong to the overlying Ventersdorp volcanic system.

Auriferous conglomerates occur at several horizons in the Witwatersrand system and also in the Black Reef series. The productive beds are, however, in the upper part of the Witwatersrand, including a thickness of about 7,000 feet of quartzites and conglomerates, among which the following are distinguished, beginning from the top: Kimberley group, Bird Reef group, Livingstone Reef group and Main Reef group. The first two are each about 500 feet thick but the conglomerates contained are of low grade, rarely exceeding $3 per ton in gold. The Main Reef group, about 90 feet thick, includes several conglomerate beds more or less persistent.

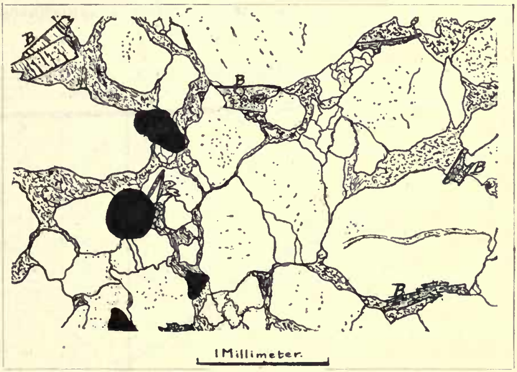

Fig. 87.—Gold-bearing conglomerate, Johannesburg, South Africa. Pebbles of quartz, crushed in places. Cement of sericite, quartz, and a little chlorite. Black areas are concretions of pyrite, replacing groundmass and quartz. B, prisms of chloritoid. *Drawn by J. D. MacKenzie.*

The usual subdivision of the Main Reef group includes from top to bottom:

South Reef (3 feet).
Bastard Reef (scattered pebbles) and quartzite (20 to 40 feet).
Main Reef Leader (2 feet).
Quartzite (2 to 20 feet).
Main Reef (4 feet).

Of these the Main Reef Leader is the most productive; the pebbles in the conglomerate are small, averaging 2 inches in diameter, and consist of well-rolled fragments of glassy quartz with fewer pebbles of more angular quartzite, banded chert, and slate. The pebbles lie in a matrix of sandy material, which has become hardened by infiltration of silica. Pyrite occurs in abundance in the cement, averaging about 3 per cent. of the rock and being present both in crystalline form and as rounded replacements after quartz probably representing two generations, both subsequent to the sedimentation. Chloritoid,[1] sericite, calcite, and graphite are other authigenetic minerals. The gold is not contained in pebbles, but only in the cement, and forms minute angular crystalline aggregates, very seldom rounded particles. It is usually closely connected with pyrite, either enclosed by it or covering the surface of pyrite aggregates. As a rule it is not visible to the naked eye.

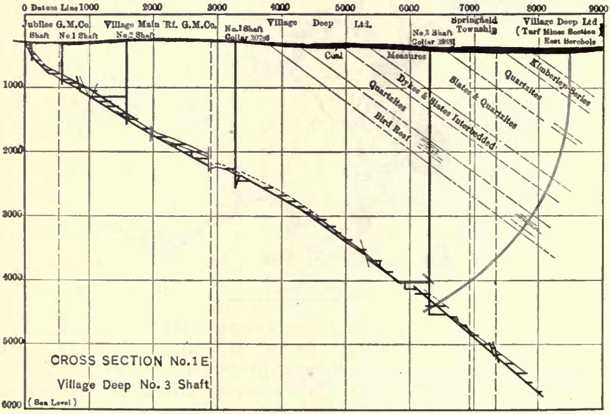

FIG. 88.—Section through Village Deep No. 3 shaft. *After H. F. Marriott.*

In spite of a long-continued discussion there is no unanimity among geologists as to the genesis of these remarkable deposits. It is evidently necessary, for a satisfactory discussion of the question, to go beyond the limits of the Johannesburg occurrences and consider the geological relations of the Transvaal and South Africa as a whole.

The first suggestion that the conglomerate may be simply an alluvial or littoral placer is refuted by the character of the gold and its close association with the pyrite. Detrital pyrite may, of

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[1] A prismatic colorless mineral usually described as chloritoid is common (Fig. 87), but its identification appears questionable.

course, occur in gravels, but there should always be some magnetite and ilmenite present. Their absence is a strong argument against the theory of direct placer deposition. It is clear that if this is a placer deposit there has been extensive recrystallization and some migration. Equally untenable is the hypothesis of F. W. Voit[1] that the gold has been brought to the surface by hot springs which discharged into the ocean.

The advocates of the placer theory, among whom are G. F. Becker, J. W. Gregory, G. A. Denny, R. B. Young and E. T. Mellor, are compelled to admit a recrystallization of the gold and a transformation of magnetite and ilmenite into pyrite.

Many geologists and engineers, impressed with the difficulties confronting the placer theory, hold that the deposits are epigenetic—that the gold and pyrite have been introduced by a post-sedimentary infiltration, perhaps after the intrusion of the diabase dikes. Such views are held by H. Louis, J. H. Hammond, R. Beck, F. H. Hatch, G. S. Corstorphine and R. B. Horwood.

Small and irregular quartz veins which in some places contain a little gold and sulphides intersect the Witwatersrand series. Whether there is any enrichment along the few diabase dikes is a disputed question. No doubt these quartz veins are related to the dikes.

E. T. Mellor has recently approached the subject from the wider geological viewpoint and his papers contain very strong arguments in favor of the original deposition of the gold in alluvial gravels. He considers the quartzite and conglomerate series as large delta deposit rather than shore gravels and shows the existence of many horizons of gold-bearing conglomerates.

Against the infiltration theory stands a long array of strong arguments: 1. The absence of channels followed by the solutions; 2. the regular distribution of the gold in the conglomerate; often it is concentrated in its lower layers; 3. the practical confinement of the gold to the conglomerates, though the quartzites are equally permeable; 4. the conglomerates were deposited in an alluvial plain skirting the deeply eroded Swaziland schists with their lenticular gold quartz veins and would thus certainly contain some gold.

The fine flake gold would be recrystallized and pressed between

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[1] F. W. Voit, Der Ursprung des Goldes in den Randconglomeraten, *Monatsber.* Deutsch. geol. Gesell., vol. 60, Nos. 5 and 7, 1908.

secondary growths of quartz grains. The original black sand would be recrystallized to pyrite by action of meteoric waters in sediments which contained sulphates and organic matter. The difficulty not yet fully explained lies in the abnormal richness and extent of the conglomerates. It is pointed out, however, that large areas of the conglomerate are practically barren. The rich beach sands of Nome, Alaska, have been cited as an analogous case but the analogy is by no means perfect. Very fine detrital gold would, of course, be expected in a delta deposit near the shore line.

Similar conglomerates of considerable geological antiquity are found in West Africa at Tarkwa and Abosso[1] and these have been worked on a fairly large scale. Instead of pyrite these contain magnetite, and ilmenite with chloritoid.

PLATINUM PLACERS[2]

It is known that platinum occurs as a primary constituent of peridotites, and specimens showing its intergrowth with olivine and chromite have been described. Almost the entire world's production is obtained from placers and 95 per cent. of it is extracted from the placers on the eastern slope of the Ural Mountains, where detrital platinum occurs in the gravels of the stream courses, which head in certain areas of peridotite and pyroxenite. It is associated with iridosmine, iridium, chromite, and often also with gold. The crude platinum forms small rounded grains, very rarely nuggets up to 20 pounds in weight, and its fineness (per thousand) ranges from 750 to 850, the remainder being iron, copper, and various metals of the platinum group, particularly iridium.

Platinum-bearing gravels occur also in Colombia, South America, in river beds and Tertiary conglomerates, and the production from them is increasing. In the United States the metal occurs in small quantities together with gold in almost all the gold-bearing districts in northern and central California

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[1] Edward Halse, *Trans.*, Fed. Inst. Min. Eng., vol. 2, 1891, p. 69.

R. Beck, Erzlagerstätten, 2, 1909, p. 200.

[2] J. F. Kemp, Geological relations and distribution of platinum and associated metals, *Bull.* 193, U. S. Geol. Survey, 1902.

Louis Duparc, Le platine et les gîtes platinifères de l'Oural. Genève, 1911.

D. T. Day, W. Lindgren and J. M. Hill, in successive issues of *Mineral Resources*, U. S. Geol. Survey.

and in southwestern Oregon, where serpentine or peridotite is found. From 400 to 700 ounces are annually recovered, chiefly from the black sands of the dredges.[1] Platinum also occurs in the beach sands of southern Oregon, together with more or less gold; a small quantity of this is usually also recovered. The Tulameen district, British Columbia, formerly yielded some production. The normal world's production of platinum before the war was about 300,000 troy ounces, but it is much less at the present time. Regarding platinum and palladium in vein deposits and in peridotite see p. 790, where the uses of the platinum metals are also described. For a long time the price of platinum was less than that of gold; a gradual rise increased the value to $20 per ounce, and in 1911 it had reached $45. Increasing scarcity forced the price up to $105 in 1917. Crude platinum with 70 per cent. to 85 per cent. Pt is sold from $30 to $60 per ounce.

CASSITERITE PLACERS[2]

The original home of cassiterite (or oxide of tin) is either in the granites, in pegmatite dikes, or in quartz veins. From any of these sources it may be set free by weathering and disintegration, and, on account of its high specific gravity, it easily becomes concentrated in gravel deposits of different types. Among the accompanying minerals tourmaline, topaz, and wolframite are the most common. Grains of metallic tin are reported to occur with cassiterite in Nigeria and Australia. Eluvial deposits immediately below the croppings are numerous and are worked on a large scale at Mount Bischoff, in Tasmania. A small deposit of this kind resting in a shallow gully immediately below a pegmatite dike was mined near Gaffney, South Carolina, in 1905. The earliest production of stream tin came from gravels below the tin-

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[1] J. M. Hill, Platinum, *Mineral Resources*, U. S. Geol. Survey, pt. 1, 1917.

James W. Neil, Recovery of platinum in gold dredging, *Min. and Sci. Press*, Dec. 8, 1917.

[2] Sydney Fawns, Tin deposits of the world, London, 1907.

H. W. Kayser and R. Provis, The Mt. Bischoff tin mine, *Proc.*, Inst. C. E. (London), vol. 123, 1896, pp. 377–387.

O. H. Van der Wyck, The occurrence of tin ore in the islands of Banca and Billiton, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, 1896, pt. 3, pp. 227–242. *Mineral Resources*, U. S. Geol. Survey, 1895.

L. C. Graton, Reconnaissance of some gold and tin deposits of the southern Appalachians, *Bull.* 293. U. S. Geol. Survey, 1906.

bearing lodes of the Erzgebirge, in Saxony, and of Cornwall, both sources now practically exhausted. Considerably more than one-half of the world's production of about 123,000 short tons of tin is still obtained from placers, mainly in the Malay Peninsula and the islands of Banca and Billiton, near Sumatra. New South Wales and Victoria furnish minor amounts and in the latter state some cassiterite is saved in working Pliocene auriferous stream channels. In this case the tin ore appears to be sparsely disseminated in granite and is liberated after its disintegration. At the Briseis mine, in Tasmania, the deposit worked consists of 14 to 45 feet of river gravel, covered by 20 to 40 feet of decomposed basalt and containing from 2 to 4 pounds of cassiterite per cubic yard.

In the United States small amounts of stream tin are recovered in Alaska near the extreme western point of the American continent, in the Black Hills of South Dakota, and in North and South Carolina.

As tin is worth from 25 to 90 cents per pound and the easily reduced cassiterite contains 78.6 per cent. of the metal, it is clear that a small quantity, say 2 pounds per cubic yard of gravel, might suffice for profitable working. Some gravels in Tasmania now worked average only 0.6 pound per cubic yard.

MONAZITE PLACERS[1]

Monazite, an anhydrous phosphate of cerium, lanthanum, and other cerium metals, usually contains also from 3 to 8 per cent. of thoria, making it valuable for the production of nitrate of thorium, which is utilized in the manufacture of incandescent gas mantles.

The mineral has a specific gravity of 5.203, a resinous luster, and a yellow to brown color; when occurring in placers it is found together with gold, zircon, magnetite, ilmenite, garnet, etc., after concentration in sluices. From its associated minerals it is cleaned in electromagnetic separators, the final product being about 90 per cent. pure. The source of the monazite is in the granites, gneisses, and pegmatites, where it occurs as a primary mineral. As its value (changing with the percentage of thoria)

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[1] J. H. Pratt and D. B. Sterrett, Monazite and monazite mining in the Carolinas, *Trans.*, Am. Inst. Min. Eng., 40, 1909, pp. 488–511.

D. B. Sterrett, *Mineral Resources*, U. S. Geol. Survey, 1906, pp. 1195–1209. W. T. Schaller, *idem*, pt. 2, 1916, pp. 223–237.

is about 8 cents per pound, monazite gravels may in places form workable deposits, especially where, as often happens, gold is present. Monazite is now obtained from marine and fluviatile placers in Brazil and India, but it is also obtained from similar deposits in North and South Carolina and has lately been found in Idaho, where a large intrusive batholith of granite or quartz monzonite evidently carries the mineral sparsely distributed throughout. The principal occurrence in Idaho is at the old placer district of the Idaho Basin. In 1910 about 100,000 pounds of monazite was mined in the United States, chiefly from placer deposits in the Carolinas. The total value is stated as $12,000. Since 1910 there has been no production in the United States, the supply being obtained from the extensive deposits in Brazil.

OTHER PLACERS

Magnetite, or "black sand," has been frequently mentioned above as a product of concentration in gravels and sands and is usually derived from the disintegration of igneous rocks. Along the beaches and the bars of some rivers it may accumulate in considerable masses—for instance, on the lower St. Lawrence River, Canada, and along the Columbia River, Oregon—but it is exceptional that such deposits have been utilized.[1] More or less ilmenite is usually mixed with the magnetite.

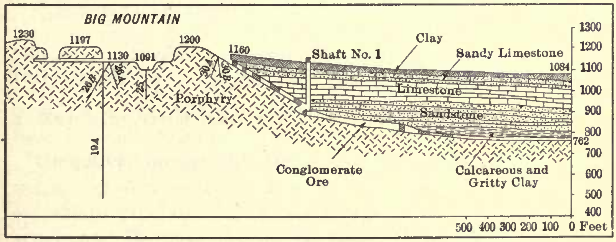

FIG. 89.—Section of Iron Mountain, Missouri, showing mining of detrital ore underneath limestone and sandstone, and of hematite ore in the porphyry. *After G. W. Crane.*

There are several examples of eluvial deposits of iron ore (magnetite, hematite, or limonite), formed below croppings of iron deposits, and also of such detrital masses in the débris slopes

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[1] The magnetite sands of Japan appear to have been rather extensively utilized; also those occurring along the coast of New Zealand.

of older formations. At Iron Mountain, Missouri,[1] Paleozoic rocks rest upon a deposit of boulders of iron ore and porphyry, which in turn lie upon pre-Cambrian porphyry. The porphyry itself also contains deposits of hematite (Fig. 89).

Similar eluvial masses of copper and lead ores are found in places. We may recall the great débris mass of chalcocite below the croppings of the Bonanza mine[2] in the Copper River region, Alaska, and galena beds on the slopes below the Elkhorn mine, Wood River, Idaho.[3]

Some placers yield precious stones. Diamonds are believed to occur as a primary mineral of some peridotites, possibly also in other rocks. Diamond placers have been worked in Brazil, India, and South Africa. In the last-named region fine stones are found in the gravel of the Vaal River, and small diamonds have lately been washed from the beach sands of Lüderitz Bay, German West Africa. In a few places sapphires, more rarely rubies (both aluminum oxide), are recovered from gravels. Along the Missouri River near Helena, Montana,[4] a bar was worked for several years for the sapphires it contained. They were plentiful, but a large proportion were of yellowish or pale blue color.

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[1] F. L. Nason, Report on the Iron Mt. sheet, Geol. Surv. Mo., vol. 9, 1912.

[2] F. H. Moffit and S. R. Capps, *Bull.* 448, U. S. Geol. Survey, 1911, p. 89.

[3] W. Lindgren, *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, p. 210.

[4] D. B. Sterrett, *Mineral Resources*, U. S. Geol. Survey, pt. 2, 1910, p. 877.

CHAPTER XVI

DEPOSITS PRODUCED BY CHEMICAL PROCESSES OF CONCENTRATION IN BODIES OF SURFACE WATER BY REACTIONS BETWEEN SOLUTIONS

LIMESTONE

Definition and Origin.—The limestones are sedimentary rocks, composed of carbonate of calcium, usually calcite, but in recent deposits also aragonite;[1] they contain minor amounts of magnesium and iron, also varying amounts of alumina and silica, and by the increase of these constituents transitions to shale or sandstone result. Phosphate of calcium and organic matter also enter into the composition of most limestones. The rocks are always crystalline, for there is no such mineral in nature as amorphous carbonate of calcium, but the grain varies between the widest limits.

When water containing bicarbonate of calcium is discharged into the ocean or bodies of fresh water the calcium carbonate is often precipitated because of changing equilibrium in the solutions.[2] This is exemplified by deposits along the shore of the Great Salt Lake in Utah. Such limestones often form "oölitic" beds of small, rounded concretions.

Generally, organic life plays a most important part in the deposition of calcite either indirectly by precipitation by ammonium carbonate generated by decaying organisms, or directly by life processes. Bacteria[3] may be the agent, very often also algæ, the latter both in the sea and in fresh water deposits. Mollusks, corals, crustacea and echinoderms segregate calcite and aragonite[4] in their shells which accumulate on the bottom at moderate depths.

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[1] J. Johnston, E. M. Merwin and E. D. Williamson, The several forms of calcium carbonate, *Am. Jour. Sci.*, 4th ser., vol. 41, 1916, pp. 473–512.

[2] J. Johnston and E. D. Williamson, The rôle of inorganic agencies in the deposition of calcium carbonate, *Jour. Geol.*, vol. 24, 1916, pp. 729–750.

[3] T. W. Vaughn, Chemical and organic deposits of the sea, *Bull.* Geol. Soc. Am., vol. 28, 1917, pp. 933–944.

G. H. Drew, Publ. 182, Carnegie Inst. Washington, 1914, pp. 1–78.

[4] Aragonite is the unstable form of calcium carbonate and always tends to change to calcite. It is most common in recent or Tertiary deposits.

Many organisms, such as sponges, secrete silica from the sea water and thus cherty deposits may be admixed with the limestones. Many limestones are almost wholly made up of shell remains but in others no trace of organic structure may be visible. Metamorphism tends to increase the grain and destroy the fossils. Evaporation of ordinary surface waters in dry climates may produce thick beds of porous limestone. This is known in Mexico as "caliche."

Calcite and aragonite are often deposited in large masses by hot springs containing bicarbonate of calcium, and such deposits may closely simulate limestones. Certain beautiful banded and translucent spring deposits are called onyx and are used for ornamental stones.

Among the many varieties of limestone the following may be mentioned:

Chalk.—This is a white, fine-grained, loosely coherent limestone of comminuted shells of mollusks and also of foraminifera. Its occurrence in the Cretaceous along the English coast is well known. Extensive beds are reported from Texas, New Mexico, Arkansas, and Kansas. Chalk is used as fertilizer, for whiting, for marking, for polishing powder, and for many other purposes. "Paris white" is a pigment made by grinding "cliffstone," a hard variety of chalk. Much of this is imported.

Lithographic Stone.—The variety of limestone used for engraving and the reproduction of colored plates is a fine-grained rock with imperfect conchoidal fracture, gray or yellowish in color, and uniform in texture. It must be porous, to absorb the grease in the printer's ink, and soft enough to work readily under the engraver's tool. Lithographic stone of good quality is difficult to find. The product from the Solenhofen quarries in Bavaria is a Jurassic limestone of unusual excellence. The material is variable in composition and its value is ascertained only by trial.[1]

Lithographic stone is reported to occur in several States of the Union, but none of it appears to be as good as Solenhofen rock. The plates used are 22 or 28 by 40 inches, and 3 inches thick. The better grades are expensive, selling at about 22 cents per pound. The best grade quarried in the United States is said to come from Brandenburg, Kentucky.[2] The demand is limited.

[1] S. J. Kübel, *Mineral Resources*, U. S. Geol. Survey, 1900, pp. 869–873.

[2] E. O. Ulrich, *Eng. and Min. Jour.*, vol. 73, 1902, p. 895.

Hydraulic Limestone.—Certain argillaceous limestones or dolomitic limestones are used for the manufacture of natural cement. Such rock, crushed and burned, hardens or "sets" when mixed with water, owing to hydration and crystallization of the silicate and aluminate of calcium formed during the burning. "Portland" cements are mixtures of limestone and argillaceous rocks, subjected to a similar process of grinding and burning.[1]

Lime.—Pure limestones are changed by burning and consequent decarbonation to quicklime which is usually shipped as "lumplime;" this "slakes" to a calcium hydroxide when mixed with water and is used, with the addition of sand, as mortar in brick constructions. The "slaking" is retarded by the presence of magnesia and argillaceous impurities.[2] Slaked lime, finely powdered is a recent product used extensively for water proofing concrete.

Uses.—While limestone is mainly a structural material it is also used as a flux in smelting operations. Only the pure varieties are acceptable, though magnesian limestones are also used in iron furnaces.

Burnt lime is probably employed for more purposes than any other natural product. It is used for the manufacture of bleaching powder, ammonia, calcium carbide, fertilizers, wood alcohol, soap, glycerine, glue, glass, pottery, paints, paper and sugar. Also for tanneries and as insecticide and fungicide.

DOLOMITE[3]

Pure dolomite contains 54.35 per cent. $CaCO_3$ and 45.65 per cent. $MgCO_3$. Beds of dolomite and dolomitic limestone are common in sedimentary deposits. They may often be distinguished by a fine-grained sugary texture, due to a development

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[1] E. F. Burchard, *Mineral Resources*, U. S. Geol. Survey, 1916, with list of literature.

[2] E. F. Burchard and W. E. Emley, The source, manufacture and use of lime, *Mineral Resources*, U. S. Geol. Survey, pt. 2, 1913, pp. 1509–1593.

[3] F. M. Van Tuyl, New points on the origin of dolomite, Am. Jour. Sci., 4th ser., vol. 42, 1916, pp. 249–260.

F. M. Van Tuyl, The origin of dolomite, *Ann. Rept.*, Iowa Geol. Survey, vol. 25, 1914, pp. 251–422.

For an extended discussion of dolomite see F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, pp. 559–571.

E. Steidtmann, Origin of dolomite, etc., *Bull.* Geol. Soc. Am., vol. 28, 1917, pp. 431–450.

of uniform rhombohedral crystals. Dolomite is somewhat harder than limestone and is insoluble in dilute hydrochloric acid. Magnesium carbonate is much less soluble in water than calcium carbonate, as shown by the fact that stalactites in magnesian limestone caves are almost wholly $CaCO_3$. Some travertines from mineral springs are rich in $MgCO_3$ and may contain up to 29 per cent. of this compound. Dolomite is doubtless deposited by direct precipitation in sea water, but much of the dolomite has been formed by alteration of the limestone by sea water, or by subsequent dolomitization by surface waters.

Deep borings in coral reefs have shown that the limestone, somewhat magnesian at the surface, passes into dolomite in depth. Certain algæ deposit much $MgCO_3$ with $CaCO_3$; some shells also contain magnesium carbonate but seldom more than 7 per cent. In warm waters the percentage of $MgCO_3$ in shells tends to increase.

Instances are known of the deposition of thin beds of pure magnesite in bodies of water.

IMPORTANCE OF CARBONATE ROCKS AS RELATED TO ORE DEPOSITS

Within the zone of oxidation the carbonate rocks are often dissolved, residual clays being then developed. Accessory constituents such as barium (and strontium), probably present in most limestones but rarely determined, or zinc and lead in the form of sulphides, or admixed phosphates may then become concentrated and acquire economic importance.

Limestone is easily silicified by waters containing silica; the silica usually appears as irregular masses of fine-grained quartz or chert. It is quite as easily dolomitized by dilute waters containing some magnesia, and this is often observed near ore deposits formed at slight or moderate depth. Limestone and dolomite, under the influence of heated waters, are subject to replacement by quartz, dolomite, barite, and fluorite or by metallic ores such as pyrite, blende, and galena. At high temperature and pressure pure limestones recrystallize to marble. Silicates, such as garnet, diopside, or wollastonite, form in argillaceous or siliceous limestone from the impurities contained or from the introduction of solutions rich in silica and iron. Lastly, the limestones are easily soluble and caves develop along fractures, forming receptacles for the deposition of ores.

CHERTS AND DIATOMACEOUS EARTH

The silica accumulated by detrital processes as sandstone and quartzite has already been mentioned. Silica may, however, also be extracted from water and deposited as a sediment by means of organisms, such as radiolarians, diatoms, and sponges. In part this silica forms cherty masses included in limestone; part is deposited as distinct beds. Diatomaceous earth[1] is a deposit formed in lakes and swamps, as well as in the sea, and, when pure, consists of the delicate tests of diatoms, a class of algæ (see Arnold and Anderson, *Bull.* 315, U. S. Geol. Survey, p. 438). Such deposits accumulate abundantly where siliceous volcanic tuffs were deposited in lakes, as occurred at many places in the Cordilleran region during the Tertiary period. Thick beds are found in the Miocene of Santa Barbara County, California. The diatomaceous earth is frequently more or less admixed with rhyolitic glass and other detritus; the tests consist of hydrated opaline silica. The earth forms light-colored beds of extremely fine texture and it finds extensive use as a polishing powder, a steam-pipe packing, and an absorbent for various liquids. It contains up to 87 per cent. SiO_2 and 5 to 9 per cent. H_2O.

SEDIMENTARY SULPHIDE DEPOSITS

As the sedimentary rocks largely consist of the detritus of the continents, it is self-evident that they may contain the metals of the rocks and ore deposits of the land areas. Iron is, of course, abundant, also in a lesser degree manganese; concretions of hydrous oxides of manganese are found in the deep sea deposits and analysis shows that these contain notable amounts of nickel, cobalt, copper, zinc, lead, and molybdenum. Many limestones have been shown to contain minute amounts of zinc, lead, and copper. River sands and gravels and even littoral ocean sands may, locally, contain some detrital pyrite, but it is extremely unlikely that a sufficient quantity of these sulphides would escape oxidation to form important deposits.

Two very important analyses of composite samples have been published recently by the U. S. Geological Survey and are given

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[1] W. C. Phalen, and F. J. Katz, *Mineral Resources*, U. S. Geol. Survey, 1908–1916, under heading "Abrasives."

For the preparation and uses, see Percy A. Boeck, *Met. and Chem. Eng.*, vol 12, 1914, pp. 109–113.

below.[1] The rarer elements determined are of special interest. There can be no doubt now that copper, lead and zinc as well as nickel and cobalt are contained in marine and fluviatile deposits but the quantities seem to average smaller than these determined in paleozoic carbonate rocks, and considerably smaller than those obtained from pre-Cambrian igneous rocks (p. 10).

ANALYSES OF SILT AND DELTA DEPOSITS

	1	2
SiO_2	46.64	69.96
Al_2O_3	14.08	10.52
Fe_2O_3	4.14	3.47
FeO	1.88	
MgO	1.95	1.41
CaO	7.20	2.17
Na_2O	2.98	1.51
K_2O	1.84	2.30
H_2O-	4.73	3.78
H_2O+	5.86	1.96
TiO_2	1.84	0.59
ZrO_2	None	0.05
CO_2	4.05	1.40
P_2O_5	0.17	0.18
SO_3	0.32	0.03
Cl	2.25	0.30
F	.....	0.07
S	0.11	0.07
BaO	0.05	0.08
SrO	0.025	Trace
MnO	0.10	0.06
V_2O_3	0.028	0.02
Cr_2O_3	0.044	0.01
MoO_3	None	....
As_2O_3	Trace	0.0004
(Ni, Co)O	0.080	0.017
CuO	0.009	0.0043
PbO	0.0004	0.0002
ZnO	0.0087	0.001
C	1.38	0.66
	101.775	100.6229
Less O	0.56	0.12
	101.215	100.5029

1. Terrigenous blue mud; composite of 52 samples. George Steiger, analyst.

2. Mississippi delta mud; composite of 235 samples. George Steiger, analyst.

Pyrite, and infrequently other sulphides, may be precipitated by chemical reactions in sediments. Beds of oölitic pyrite are known (p. 269); iron disulphide is formed in places in bogs

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[1] C. E. Siebenthal, Zinc and lead deposits of the Joplin region, *Bull.* 606, U. S. Geol. Survey, 1915, p. 72. With extensive discussion of results.

and streams or in oceanic sediments where hydrogen sulphide developed by decaying organic matter reacts on the sulphates of iron. If these sediments are brought to the surface by orogenic movements and slightly metamorphosed, the sulphide, originally in fine dissemination, may recrystallize in more prominent form. As a matter of fact, the deep sea muds thus far analyzed contain little or no pyrite. In the special case of the Black Sea, often quoted of late from N. Androussof's description,[1] microörganisms assist in liberating hydrogen sulphide, part of which, by reaction with iron from the sediments, develops pyrite.

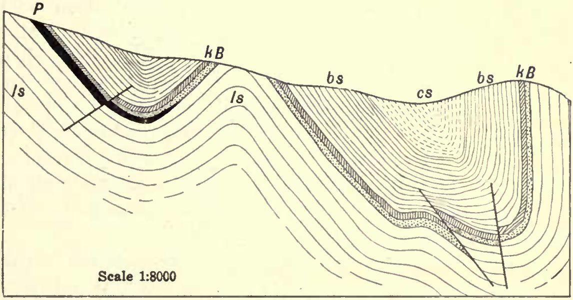

Fig. 90.—Vertical section through the pyritic deposit at Meggen, Germany. *P*, Pyrite; *B*, barite; *k*, limestone; *bs*, *cs*, *ls*, Devonian slates. *After Strauss.*

Although pyritic clays are abundant in the unmetamorphosed sedimentary formations, there is little evidence of extensive sedimentary beds of pyrite.

The deposit at Meggen, in Westphalia, is often referred to as tending to prove the existence of sedimentary beds of pyrite. Oölitic pyrite with barite and zinc blende occurs here in Devonian beds and is worked on a fairly extensive scale the bed being from 12 to 20 feet thick (Fig. 90). Alfred Bergeat[2] appears to have demonstrated the sedimentary origin of the oölitic pyrite. The barite and the zinc blende are somewhat later than the pyrite and their origin is still doubtful. Bergeat holds that all these minerals form a local marine deposit.

Barite when appearing in sandstone is probably deposited

[1] N. Androussof, Guide des excurs. du VII Cong. géol. int., 29, 1907, p. 6.

[2] A. Bergeat, *Zeitschr. prakt. Geol.*, vol. 22, 1914, pp. 237–249.

by hot springs; it is difficult to conceive its origin from sea water. For barite and manganese carbonates as marine shore deposits see p. 269.

Evidence is scant as to the sedimentary deposition on a large scale of sulphides other than pyrite or marcasite. "The Kupferschiefer" of Mansfeld, in which the sulphides may be of syngenetic origin, will be described elsewhere (p. 413).

SEDIMENTARY IRON ORES

It is conceded that iron ores, such as magnetite, can be deposited by mechanical concentration as placers along rivers or the seashore (p. 245), or again we may easily conceive hematite or limonite derived from deep decay of rocks along the littoral, or from the oxidation of pyrite deposits, as at Rio Tinto, Spain, swept out into the sea and deposited close to the shore. Iron ores are also formed by chemical reactions in bodies of water, and these yield a notable proportion of the iron production of the world. In the latter cases the iron has been supplied from the land areas in form of solutions. In many instances both dissolved iron salts and detrital minerals of iron contribute to the genesis of the deposits.

The surface waters extract iron from ferromagnesian silicates as well as from oxides or other minerals; this extraction proceeds most energetically in regions covered by a deep mantle of decayed rock. Both iron and manganese are contained in springs and streams. An example of such spring water, rising underneath a deposit of bog iron ore in Holland, is quoted by Clarke:

ANALYSIS OF SPRING WATER AT EDERVEEN, NETHERLANDS [1]

(Analyst, G. Moll van Charante)

(Parts per million)

Ca	107.6	Al_2O_3	3.3
Mg	5.6	Cl	15.2
Fe	19.6	H_3PO_4	10.9
Mn	11.4	SO_4	0.9
K	0.9	CO_3	207.6
Na	10.0	SiO_2	18.0
		Organic	56.0
			467.0

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[1] Cited by F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 530.

A larger part of the iron dissolved by the surface water is precipitated after a short journey,[1] but some of it is carried down by the streams to lakes and seas, in which it then may be deposited on an extensive scale. The sea water contains about 0.6 milligram per liter of iron and probably more at some places near the shore.

LIMONITES IN SWAMPS AND LAKES (BOG IRON ORES)

Occurrence.—The bog iron ores are found in swamps, lakes, or even in sluggish water courses, and they are especially abundant in the recently glaciated regions of northern Europe, Asia, and North America. They consist of dark-brown, rough and cellular masses or loose particles, sometimes oölitic in structure and then designated as "shot ore," and form a layer of varying thickness at the bottom of the swamp or lake. Plants and roots may be replaced by limonite. Such ores are usually mined by means of primitive dredges or scoops.

The ore occurs mainly in shallow waters along the shore, to a depth of about 12 feet. After removal a new layer is formed within a few years; according to A. Geikie[2] several inches of limonite accumulated in 26 years in a Swedish deposit. The rate would naturally be subject to great variations according to local conditions.

The bog iron ores are now of slight importance to the mining industry, but the easily traceable processes of their formation give us a most welcome key to the origin of other and more obscure deposits.

Composition.—These ores are always mixed with sand and clay and rarely contain as much as 50 per cent. of iron. The principal mineral contained is limonite, but carbonate of iron is commonly present, also phosphate as vivianite; soluble silica is sometimes recorded. In some low-grade ores from the Netherlands, the analyses of which are quoted by Clarke,[3] there is much more ferrous carbonate than limonite. Varying quantities of manganese are present in ores from Sweden, Finland, and Holland. The Swedish ore contains traces of vanadium, molybdenum, copper, lead, zinc, arsenic, nickel, and cobalt. All bog iron ores contain phosphorus, but there is rarely much sulphur.

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[1] The clogging of water supply pipes by hydroxides of these metals is a common occurrence.

[2] A. Geikie, Text-book of geology, 4th ed., 1903, p. 187.

[3] F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 530.

According to Svanberg, cited by Zirkel,[1] the average of 30 analyses of Swedish bog iron ores gave:

Fe_2O_3	62.57	MgO	0.19
Mn_2O_3	5.58	P_2O_5	0.48
SiO_2	12.64	SO_3	0.07
Al_2O_3	3.58	Ignition	13.53
CaO	1.37		
		Total	100.01

Origin.[2]—The agents by which iron is carried into solution are (1) carbon dioxide from the air and decomposing organisms; (2) sulphuric acid from the weathering of pyrite, and (3) organic acids derived from decomposing vegetable matter. In the absence of air ferric oxide is reduced to the ferrous state and forms soluble double salts with ammonia and humic acid.

Precipitation is effected in bicarbonate solutions by the escape of carbon dioxide in the air or through its absorption by plant cells. The ferrous carbonate is easily oxidized to ferric hydroxides. In the presence of much organic matter ferrous carbonate remains in the precipitate.

From ferrous sulphate solution iron is precipitated as limonite by oxidation and hydrolysis, or by reaction with calcium carbonate solution, in which case siderite and gypsum will result, the former oxidizing to limonite, or the iron may be precipitated by ammonium humate, always present in swamp waters, or finally by soluble calcium phosphate, in which case vivianite or other iron phosphates result. Less commonly the iron is precipitated as pyrite by alkaline sulphides or hydrogen sulphide.

From soluble humates iron is also precipitated by organisms, called iron bacteria, which take up these humates, as well as ferrous carbonate, and coat their cell walls with the segregated limonite, but regarding the real importance of this process we have few data.

In these, as in so many other surface reactions, the ferric

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[1] F. Zirkel, Lehrbuch der Petrographie, vol. 3, 1894, p. 574.

[2] R. Beck, Die Lehre von den Erzlagerstätten, 3d ed., vol. 2, 1909, p. 397.

F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 529.

Ossian Aschan, *Zeitschr. prakt. Geol.*, 1907, pp. 56–62.

C. R. Van Hise, Metamorphism, *Mon.* 47, U. S. Geol. Survey, 1904, p. 826.

J. M. van Bemmelen, *Zeitschr. anorg. Chemie*, vol. 22, 1906, p. 313.

J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1894, p. 30, and 1895, p. 38.

F. M. Stapff, *Zeitschr.* Deutsch. geol. Gesell., vol. 18, 1866, p. 86.

hydroxides are probably precipitated as colloidal complexes of indefinite composition, or "gels," which in time tend to change to crystalline bodies. Much of the ferric hydroxide is doubtless transported for considerable distance in colloid form. Five species of ferric hydroxide are recognized. Arranged by increasing water they are:

Turgite	$2\ Fe_2O_3.H_2O$	94.6	per cent.	Fe_2O_3
Goethite	$2\ Fe_2O_3.2H_2O$	89.9	"	"
Limonite	$2\ Fe_2O_3.3H_2O$	85.5	"	"
Xanthosiderite	$2\ Fe_2O_3.4H_2O$	81.6	"	"
Limnite	$2\ Fe_2O_3,6H_2O$	74.7	"	"

Both göthite and turgite are red and may be mistaken for hematite. Limonite and göthite, and probably also the other compounds, soon acquire crystalline properties and then show strong double refraction and fibrous texture. There are several complex ferric silicates and sulphates, which look somewhat like limonite.

Examples.—Many occurrences of bog iron ores are known in the United States. At Radnor and Drummondville, Three Rivers district, Quebec, the occurrences are extensive. The ores contain 0.3 per cent. phosphorus and less than 0.1 per cent. sulphur. This iron ore was utilized until 1911, being dug in the swamps or dredged in the lakes; 23,000 short tons were mined in 1907, but in 1911 the operations ceased. The production in this district began in 1733.

One of the most famous deposits formerly mined is at Katahdin, Maine. Small deposits are found at very many places in New England and have been worked on a small scale. Near Portland, Oregon, at the Prosser mine, limonite ore was found in the surficial hollows of a basalt flow, covered by a later flow of the same rock; it was 6 to 15 feet thick and contained roots and trunks of trees.[1] The earlier basalt was rich in iron and its decomposition furnished the iron to the swamps which covered its surface.

THE SIDERITES OF MARINE AND BRACKISH-WATER STRATA

Occurrence.—Siderite ($FeCO_3$) is an iron ore of some importance, both in epigenetic and syngenetic deposits. It occurs in fissure veins and as a replacement of limestone, but is also found in the sedimentary rocks as a product of the sedimentary proc-

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[1] J. F. Kemp, Ore deposits of the United States and Canada, 1900, p. 92.

esses. The sedimentary siderite ores are called clay ironstone, spherosiderite, or black band. A dense or fine-grained concretionary structure is characteristic of the "clay ironstone" occurring in clays or shales and these concretions, more or less admixed with clay and sand and often inclosing vegetable remains. are found abundantly at certain horizons. The variety called "black band" forms continuous beds of dark-colored, compact appearance, in the shales of the coal measures, and often directly underneath or above the coal beds.

These ores contain less than 48 per cent. of iron and must be calcined before smelting. Both sulphur and phosphorus are present, sometimes in considerable quantities.

Marcasite, pyrite, arsenopyrite, millerite, galena, blende, and chalcopyrite are sometimes found along cracks in the concretions of siderite, indicating that the iron solutions carried small amounts of the less common metals, probably as sulphates. After the deposition of the siderite these metals were leached and redeposited as sulphides along available openings. An analysis of siderite ore from Maryland[1] showed 36.05 per cent. Fe, 13.53 per cent. SiO_2, 6.47 per cent. Al_2O_3, 0.94 per cent. Mn, 0.08 per cent. P, and 0.42 per cent. S.

The economic importance of these ores, formerly great, is now small. Near the surface they are sometimes changed to limonite.

The origin of sedimentary siderites is explained along the same lines as that of the bog iron ores. Solutions of ferrous bicarbonate were supplied to the marshes along the sea coast or to the shallow sea where organic matter was abundant. Precipitation of the normal insoluble carbonate took place through absorption of the solvent CO_2 by vegetation. Free oxygen was absent, for otherwise the carbonate would have been transformed into limonite. Even if the iron had originally been deposited as limonite a reduction and carbonation to siderite may have been effected by the limonitic precipitate being covered by mud containing organic matter.

The concretionary ores are not the products of primary precipitation, but are, probably in all cases, segregated into nodular form by the action of percolating solutions around a suitable nucleus while the sediments were still soft.

[1] J. T. Singewald, *Econ. Geol.*, vol. 4, 1909, pp. 530–543.

J. T. Singewald, Report on the iron ores of Maryland, Maryland Geol. and Econ. Survey, vol. 9, pt. 3, 1911.

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Examples.—In the United States sedimentary siderites are known in Pennsylvania, Ohio, West Virginia, Maryland, and Kentucky. Their present industrial importance is slight, but they were formerly mined on a more extensive scale. The production in 1916 was only 1,800 long tons, chiefly from Ohio.

In Pennsylvania and adjacent States the upper barren Coal Measures contain abundant nodules of siderite in the shales and sandstones, but no valuable deposits. In the upper productive Coal Measures, or Monongahela River series, black band ore occurs, for instance, just below or above the Pittsburg coal bed. In the lower Coal Measures the siderites are especially abundant; in Ohio 12 horizons of black bands and concretionary ores are distinguished by Orton.[1]

Siderite ores also occur in the Tertiary Claiborne formation of Mississippi.

The black bands are common in Germany, but are not mined extensively. They were formerly actively worked in Westphalia and near Saarbrücken, where the ore formed flat lenticular masses as much as 1½ meters thick and sometimes several hundred meters in extent.

In England the black bands were formerly of the highest importance and 40 years ago furnished four-fifths of the total iron output. They are now mined only in North Staffordshire and in Scotland. In Wales the black bands occur in the lower Coal Measures. Kendall[2] enumerates 75 horizons of siderite ore.

In Scotland (Ayrshire) the black bands occur both in Coal Measures and in Carboniferous limestone. The ores contain 25 to 40 per cent. Fe, and occur as thin strata, 1½ feet or less thick; several of them are usually close together.

THE JURASSIC SIDERITES OF ENGLAND

The carbonate ores of the Jurassic "oölite" in England have a much greater importance than formerly,[3] the mine production

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[1] Ohio Geol. Survey, vol. 5, 1884, p. 378.

[2] J. D. Kendall, The iron ores of Great Britain and Ireland, London, 1893, pp. 145–199.

[3] Henry Louis, The iron ore resources of the world, Int. Geol. Congress, Stockholm, 1910.

W. G. Fearnsides, British iron ore resources, *Min. Mag.*, London, Nov., 1917, pp. 241–243.

W. Barnes, Mining iron ore in the Midlands, *idem*, March, 1918, pp. 120–126.

being about 11,000,000 metric tons, out of a total output of 15,000,000 tons. While the ores are of low grade they are cheaply mined, largly in open cuts. The largest yield comes from the Cleveland Hills, in the Yorkshire district. The ores form three or four beds, in the shales and sandstones of the Lower Oölite, Upper Lias, and Middle Lias; the thickest bed attains 13 feet with several minor clay streaks. The ore is changed to limonite near the surface, but the bluish green unaltered rock is composed largely of oölitic siderite; a little glauconite is present. Its percentage composition is approximately as follows: SiO_2, 10 to 20; FeO, 40; Fe_2O_3, 1.4; CaO, 1.5; CO_2, 25; P_2O_5, 0.5 to 2. There is little sulphur and the metallic iron varies between 29 and 35 per cent. A little magnetite is reported in the ore.

THE OÖLITIC MARINE LIMONITES AND HEMATITES

The oölites (name derived from the semblance to fish-roe) consist of small rounded grains of concretionary origin, each grain often being formed around a small sand-grain or around a small fossil fragment. They are found in shallow water near the shore, where the action of waves and currents is strong. Oölitic limestones are common occurrences in some sedimentary series. The oölitic iron ores consist of limonite, of hematite, of siderite or of iron silicates. Frequently all these occur together. The concretions are cemented by calcite or siderite or more commonly by an argillaceous substance.

THE OÖLITIC LIMONITES

Occurrence.—The oölitic limonites form well-defined and extensive beds in purely sedimentary series of sandstones, shales, and marls. Though several ore beds are usually present in each district they are not always persistent, but may thin out, other beds appearing at different levels. The ores have no relation to volcanism, though in many cases the decay of volcanic rocks may have supplied the iron. Though not particularly characteristic of any one formation the ores are most abundant in Jurassic strata. The percentage of iron is low and that of phosphorus high; favorable features are the presence of calcium carbonate, which makes the ores self-fluxing, and the great extent of the beds. The great iron industries of Germany and France are largely dependent upon the oölitic limonites.

Examples.—The so-called "minettes,"[1] or öolitic limonites of the German and French Lorraine and of Luxembourg, are of the highest importance as present and future resources of European iron. In France there are at least 50 mines with an annual ore production of nearly 20,000,000 metric tons (1913). The proved reserves are estimated at 3,000 million tons. In German Lorraine, now ceded to France, the production attains similar figures and the estimated reserves are over 2,000 million tons. Dipping gently westward the strata attain a depth of 3,000 feet or more in France. The present mining is done at a depth of 700 feet or less, and in part by tunneling or open cuts (Fig. 91),

The ores lie in the Middle (Dogger) part of the Jurassic systems and occur with shales, sandstones, and marls as distinct bed. within a vertical distance of 75 to 150 feet. The strata are not absolutely persistent at the same level, but are local accumulations, thinning out in lenticular manner. The several beds known are of different thickness, the maximum being 15 feet. A low percentage of iron, varying from 31 to 40, is characteristic, likewise a high percentage of phosphorus, varying from 1.6 to 1.8, the latter making the ores available for the basic process. From 5 to 12 per cent. CaO and from 7 to 33 per cent. SiO_2 are present. The ores are earthy and soft and are of brown, gray, or yellow

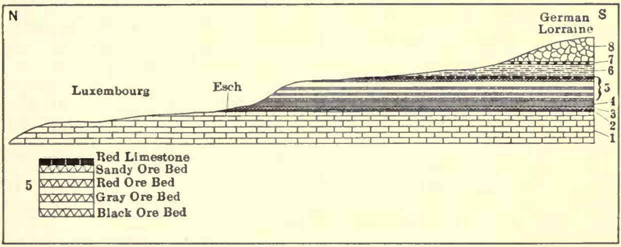

FIG. 91.—Section through the minette measures at Esch; 8, Calcareous layer with *Harpoceras humphriesianum;* 7, calcareous layer with *Harpoceras sowerbyi;* 6, marl with *Harpoceras murchisonæ;* 5, the minette measure group (see legend); 4, sandstone with *Trigonia navis;* 2 and 3, upper and lower clays with *Harpoceras striatulum;* 1, Lias (micaceous marl). *After W. Branco.*

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[1] P. M. Nicou, in "Iron ore resources of the world," Int. Geol. Congress, Stockholm, 1910.

L. van Werveke, *Zeitschr. prakt. Geol.*, 1895, p. 497, 1901, pp. 396–403.

tints. Limonite forms the bulk of the ore, but there is always calcite and some siderite, often also a little secondary magnetite, and more rarely grains of pyrite, zinc blende, galena, and chalcopyrite. The small concretions of the size of a pin-head, or a little larger, consist of limonite but, like the Clinton oölites of hematite, have a skeleton of silica, which, according to van Werveke, points to a probable derivation by alteration from glauconite. The cement consists of silica, lime, or clay shale, and grains of glauconite, a ferri-potassic silicate, occur in it.

Less important öolitic ores occur in other parts of Europe, likewise in the Jurassic system.

Cretaceous oölites in which the limonite is probably derived from glauconite or siderite have been described from Texas.[1]

Origin.—The origin of these limonites is a disputed question. As already indicated some observers doubt the direct deposition of limonite in the sea water, but hold that the mineral resulted from the oxidation of oölitic siderite or glauconite.

In a recent detailed monograph L. Cayeux[2] emphasizes the peculiar fact that the limonitic oölites of France are confined to the older, pre-Cretaceous formations, while in the Cretaceous or later beds the glauconites predominate. He believes that the older oölites are in all cases derived from siderite by replacement and oxidation.

THE MARINE OÖLITIC SILICATE ORES

A number of silicates of iron are distinctly sedimentary products and common in many waterlaid series of rocks; the most important are glauconite (greensand), chamosite, thuringite, and greenalite. Glauconite seems especially abundant in the Cretaceous, chamosite and thuringite in the Silurian, and greenalite in the Algonkian, but none of them are confined to any definite horizon. Their composition is uncertain and variable, the glauconite $KFeSi_2O_6.H_2O$ alone being distinguished by a large percentage of potash.

Glauconite[3] forms in marine deposits on the present sea

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[1] R. A. F. Penrose, *Bull.* Geol. Soc. Am., vol. 3, 1891, p. 47.

[2] L. Cayeux, Les minérais de fer oölithique de France, Paris, 1909.

[3] C. W. von Gümbel, *Sitzber.* Akad. München, vol. 16, 1886, p. 417; vol. 26, 1896, p. 545.

L. Cayeux, Contribution étude microgr. des terr. séd., Lille, 1897.

W. A. Caspari, Contributions to the Chemistry of the Marine Glauconite, Proc. Edinb. Roy. Soc., vol. 30, 1909, p. 364.

bottom and also occurs scattered in marine sands of older formations from the Cambrian to the present time, sometimes so abundantly that the rocks are termed greensands. The Cretaceous greensands of New Jersey form a good example of rocks containing abundant glauconite; they are rich in both phosphorus and potassium. The mineral occurs as dark-green granules, often in the interior of shells.

According to Murray and Renard, glauconite is formed just beyond the limits of wave and current action, where the muddy deposits begin. Organic matter is believed to reduce the iron in the mud to sulphide, which later oxidizes to limonite; at the same time colloidal silica is set free and the colloidal limonite absorbs this as well as potash, forming ferric silicate.[1]

APPROXIMATE COMPOSITION OF SEDIMENTARY IRON SILICATES

	Glauconite	Chamosite	Thuringite	Greenalite
SiO_2	53	29	24	30
Al_2O_3	10	13	17	..
Fe_2O_3	21	6	15	35
FeO	2	42	33	26
CaO	1	..	..	..
MgO	3	..	..	..
K_2O	4	..	..	..
H_2O	6	10	11	9

In part, however, as shown by Cayeux, the glauconite has been formed somewhat later than the original deposition of the beds and without the intervention of organic matter.

Greenalite occurs abundantly, according to Leith[2] in the Algonkian ferruginous cherts of the Mesabi and other iron districts; it was formerly mistaken for glauconite. It is believed to be a marine deposit.

Neither glauconite nor greenalite rocks form iron ores, but the former may be transformed by alteration into limonite, and, according to Van Hise and Leith, the greenalite is the source from which the hematites of the Lake Superior region were in part derived.

Chamosite and thuringite[3] form the principal ore minerals in

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[1] F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 516. The extensive literature is here summarized.

[2] C. K. Leith, *Mon.* 43, U. S. Geol. Survey, 1903, pp. 237–279.

[3] A. W. Stelzner and A. Bergeat, Die Erzlagerstätten, 1, 1904, p. 201. E. R. Zalinski, Neues Jahrb. 1904; Beil. B., 19, pp. 40–84.

a number of interesting deposits in Thuringia (Germany) and Bohemia, which formerly were mined extensively and which are still being mined on a small scale in the latter region. These silicates form oölitic grains in slightly metamorphosed clay slates of the Lower Silurian. The beds still retain fossils. In Germany the ores occur as lenticular beds as much as 7 feet thick. In Bohemia the Silurian series consists of slates, graywacke, and diabase tuffs. These contain beds of oölitic hematite, one bed being 16 feet thick, while other beds consist of oölitic chamosite of considerable thickness. The latter show a groundmass of siderite or chamosite, in which are embedded oölites of dark-gray chamosite. The ores are rich in phosphorus and also carry a little magnetite.

Many believe that the iron is derived from the decomposition of the associated diabase tuffs. Be that as it may, these iron ores are certainly of sedimentary origin.

THE MARINE OÖLITIC HEMATITE ORES

Occurrence.—Oölitic hematite ores of undoubted sedimentary origin are common in many parts of the world, as, for instance, Germany, France, and Bohemia. They are usually associated with Paleozoic rocks, but appear to be lacking in Mesozoic and Tertiary sediments. Siderite and calcite usually accompany them. Rarely, if ever, do they contain magnetite or metallic sulphides. Differing opinions are expressed as to their origin; they have been explained as replacements of limestone or of siderite, or again as primary sediments, the tendency in the United States being in favor of the latter theory of origin.

The Clinton Ores.[1]—The most important oölitic ores in the United States are those of the Clinton formation in the Appalachian States; they persist with remarkable regularity wherever

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[1] C. H. Smyth, Jr., On the Clinton iron ore *Am. Jour. Sci.*, 3d ser., vol. 43, 1892, p. 487.

E. F. Burchard, The Clinton iron ore deposits of Alabama, *Trans.*, Am. Inst. Min. Eng., vol. 39, 1908, pp. 997–1055.

E. F. Burchard, The red iron ores of East Tenn., *Bull.* 16, Geol. Survey Tennessee, 1913.

Burchard, Butts, and Eckel, The Birmingham district, Alabama, *Bull.* 400, U. S. Geol. Survey, 1910.

D. H. Newland and C. A. Hartnagel, Iron ores of the Clinton formation, *Bull.* 123, New York State Mus., 1908.

this formation appears. The Clinton (Silurian) lies between the Trenton limestone and the Devonian shale, and it invariably contains one or several beds of hematite ore alternating with limestone and shale. The succession of sedimentary rocks in the Birmingham district is as follows. In a general way the section applies to the entire southern Appalachian region.

	Feet
Carboniferous:	
Pennsylvanian: Pottsville formation ("Coal Measures")	2,600 to 7,000
Unconformity.	
Mississippian:	
Parkwood formation	0 to 2,000
Pennington shale (30–300 feet) Bangor limestone (670 feet) } Floyd shale	1,000 ±
Fort Payne chert	200 to 250
Unconformity.	
Devonian:	
Chattanooga shale Frog Mountain sandstone	1 to 25
Unconformity.	
Silurian: Clinton (Rockwood) formation	250 to 500
Unconformity.	
Ordovician: Chickamauga (Pelham) limestone	200 to 1,000
Unconformity.	
Cambro-Ordovician: Knox dolomite	3,300
Cambrian:	
Conasauga (Coosa) limestone	1,000+
Rome (Montevallo) shale (great thickness).	

The Clinton ores extend from western New York, through Pennsylvania, Virginia, West Virginia, Kentucky, Tennessee, and northwestern Georgia, into Alabama, where, near Birmingham, they attain their greatest development. The ores constitute beds or lenses at various horizons in the Clinton formation, which forms a striking unit of red shallow water deposits underlain and covered disconformably by great thicknesses of limestone of the Cambrian and Mississippian ages. Thin beds of ferruginous sandstone, shale and oölitic hematite make up the formation, with frequent cross bedding and some conglomerates. The ores contain calcite and in some places show gradual transition to limestones.

The average thickness of the ore beds is only two or three feet but in Alabama they reach 20 feet of merchantable ore with occasional thin shale or sandstone partings. Single ore beds may extend for many miles. In the Birmingham district the Clinton beds outcrop on the east flank of an anticline and they can be traced continuously northward into Tennessee, but the actively working mines extend only for 15 miles along the outcrop. Good ore beds have been found by drilling for several miles eastward but toward the west the formation becomes more calcareous.

Four beds are known within 80 feet in the upper part of the formation two of which are worked, with a thickness of from 9 to 20 feet. The iron ores are generally sharply bounded by shale or sandstone; in places they form transitions into ferruginous sandstone.

An important iron industry is based upon the deposits in Alabama and the annual production of ore has now attained about 5,000,000 long tons, or about 8 per cent. of the total output of iron ore in the United States. Mining has been carried 4,000 feet on the dip in some of the properties, and entirely similar ore has been shown to exist by borings at a vertical depth of 2,000 feet, 2 miles eastward from the outcrop. Large reserves of ore are available in this district.

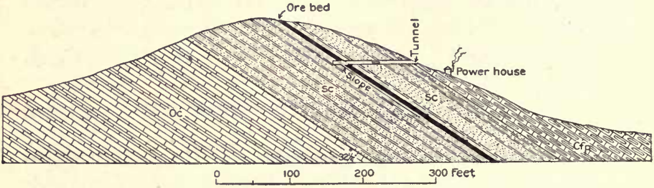

Fig. 92.—Section showing Clinton iron ores, Birmingham, Alabama; *Sc*, Clinton (Rockwood) formation, Silurian *Oc*, Chickmauga (Pelham) limestone, Ordovician; *Cfp*, Fort Payne chert, Mississippian. *After E. F. Burchard.*

Clinton ores are also mined north of Alabama in Tennessee though the operations are generally confined to the enriched surface ore. North of Tennessee the mining is profitable in few places.

There are several types of Clinton ores; most of them are fairly rich in calcium carbonate.

One common type is a fine-grained pebbly conglomerate or sandstone, each pebble or grain coated with hematite and the rock cemented with that mineral and with calcite. Another type consists largely of fragments of bryozoa, shells, trilobites, etc., partly coated or replaced by ferric oxide, besides an abundance of oölitic grains, usually with a grain of sand as the center. Still another type consists entirely of oölites of hematite in calcite matrix, averaging 1 or 2 millimeters in diameters. A fourth type shows small flattened hematite concretions with fragments of fossils changed to hematite; this is the "flax seed ore" which is very common at Birmingham. There is very little siderite or chlorite. Fragments of quartz and other minerals are common. The beds vary along the strike in their calcareous or siliceous admixtures. Phosphorus is present above the Bessemer limit that is above 0.05 per cent.

At the surface and down to a depth of about 200 feet the calcium carbonate is in part dissolved and the ore correspondingly enriched. Such ore is called "soft," in contrast to the unaltered or "hard" variety. The poorest ores used carry 25–30 per cent. iron.

ANALYSES OF CLINTON ORES

(E. C. Harder, *Mineral Resources*, U. S. Geol. Survey, 1908)

	Hard ore	Soft ore
Fe	37.00	50.44
SiO_2	7.14	12.10
Al_2O_3	3.81	6.06
CaO	19.20	4.65
Mn	0.23	0.21
S	0.08	0.07
P	0.30	0.46

The origin of the Clinton ores is a much discussed subject. The principal views demand either a direct sedimentary origin or a derivation by replacement of limestone. The latter explanation is supported by Rutledge,[1] who states that progressive steps in the transformation of limestone to ore may be followed in the field, in thin sections, and in analyses. In view of the constant character of the ore at great depth it is clear that if replacement has occurred at a comparatively late date it has at least not proceeded from the surface.

The theory of direct sedimentation is held by C. H. Smyth, who contributed a notable paper to the question of origin. Similar views are advocated by Newlands, Eckel, and Burchard. Smyth thinks that the iron was carried out into shallow marine basins and was there slowly oxidized and precipitated mechanically

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[1] J. J. Rutledge, The Clinton iron ore deposits of Stone Valley, Pennsylvania, *Trans.*, Am. Inst. Min. Eng., vol. 39, 1908, p. 1057.

around the shells or replaced them. S. W. McCallie believes that the original ore was glauconite or greenalite, citing as evidence the delicate skeleton of silica remaining when the oölite is dissolved in acid.

The Brazilian Hematites.—In the pre-Cambrian metamorphosed sediments of Minas Geraes in Brazil[1] there are thick beds of rich hematite in a formation of ferruginous sandstone (itabirite) underlain by heavy quartzite. The origin of this undoubtedly sedimentary hematite, which as yet has not been mined, is in doubt. There is no oölitic structure, nor are there fossils. Harder and Chamberlin state that "not having much confidence in the hypothesis that the iron oxide was precipitated directly from sea water by ordinary chemical means we prefer to turn to the iron bacteria as perhaps forming a better hypothesis."

The Oölitic Hematite-chamosite-siderite Ores.—Ores containing hematite, chamosite and siderite have been described from several places[2] and are in all cases of marine shallow water origin. The description by A. O. Hayes of the Wabana ores in Newfoundland is of particular interest. The ores occur in the upper 1,000 feet of flat dipping Ordovician sandstone and shale and contain several workable beds from 10 to 30 feet in thickness, one of which has been mined for a distance of one and one-half miles under the sea. The ores look like hematitic oölites and contain some fragments of marine shells but there is little calcite and no limestone. In average there is, in per cent., 50–70 hematite, 15–25 chamosite, 0–50 siderite, 0–1 calcite and 1–10 quartz. The oölites consist often of concentric shells of hematite and chamosite such as shown in Fig. 93, and are frequently embedded in a matrix of siderite. The hematite concretions, upon treatment with HCl, yield a residual skeleton of silica.

It is shown that borings of algæ penetrate both oölites and

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[1] E. C. Harder and R. T. Chamberlin, The geology of Central Minas Geraes, Brazil, *Jour. Geol.*, vol. 23, 1915, Nos. 4 and 5.

[2] L. Cayeux, Les minerais de fer oölithique de France, Ministère des Travaux publiques, Paris, 1907.

W. T. Dorpinghaus, Erzlagerstätten vom Chamosittypus . . . in der nordspanischen Provinz Leon, Archiv für Lagerstätten-forschung, Berlin, 1914.

A. O. Hayes, The Wabana iron ore of Newfoundland, *Mem.* 78, Geol. Survey Canada, 1915.

matrix and that thus the ore was practically in its present condition when covered by later sediments. Oxygen given off by these algæ may have caused oxidation of chamosite to hematite. Direct precipitation of all three iron minerals is, therefore, advocated, though siderite is believed to be the latest and may replace chamosite.

Of exceptional interest are thin beds of pyritic oölite above the "Dominion" bed. This contains graptolites, and the small pyrite concretions lie in an argillaceous matrix with some crystalline quartz.

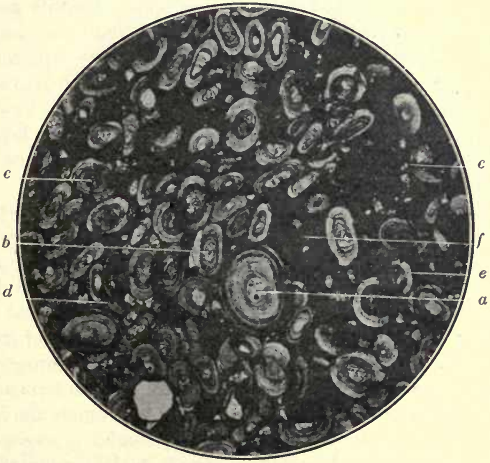

FIG. 93.—Ore from Silurian beds at La Ferrière-aux-Étangs, France. Magnified 22 diameters. The oölites are chlorite with a kernel of siderite; the fine-grained cement is chlorite and siderite. *a*, Oölite of chlorite, in center of lighter color; partly converted into hematite on the outside. *b*, Nucleus of corroded pure siderite. *c*, Same of yellow, altered siderite. *d*, Grains of siderite in the cement. *e*, Chloritic oölite, partly crushed and invaded by cement. *f*, Blackish cement of chlorite and siderite. *After L. Cayeux.*

REVIEW OF THE SEDIMENTARY IRON ORES

The descriptions given above show that in marshes, lakes and rivers the hydroxides of iron, mainly limonite, are deposited, and that smaller quantities of ferrous carbonate (siderite), iron disulphide and iron phosphates may be precipitated.

Regarding the marine ores, it is certain that glauconite and allied iron silicates are deposited in the sea and that under special reducing conditions siderite and iron disulphide may also form. The probability is also very strong that hematite is developed, in part from oxidation of siderite and glauconite, in part by detrital processes. Whether limonite is ever formed in sea water is much more doubtful for the salt solutions have a strong dehydrating effect.[1] It is more likely that the so-called marine limonites are products of oxidation of siderite and iron silicates.

The marine iron ores are all shallow water deposits and the frequent oölitic structure[2] is in part at least due to accompanying action of waves and currents. Many of the replacements observed have certainly occurred immediately after deposition. Some geologists like Cayeux hold that the ore was a limestone of organic origin which has been later transformed into hematite and siderite by successive replacements but there seems to be little to support this view.

The part played by micro-organisms is as yet difficult to evaluate. It seems certain that many of the blue-green algæ develop oxygen in their life processes which would of course promote oxidation; it is also known that sea water at all depths contains air enriched in oxygen. It is also certain that some bacteria of the Crenothrix type[3] segregate hydroxide of iron by oxidation of dilute solutions of $FeCO_3$, but this process can probably not go on in sea water. Certain other bacteria according to Drew of the de-nitrifying type seem to promote the formation of calcareous oölites in the sea, and similar processes may possibly under favorable conditions result in the precipitation of siderite.

That either siderite or pyrite can be deposited in large bodies in the open sea must be considered very unlikely.

Wherever iron disulphide is formed reducing conditions pre-

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[1] W. Spring, Neues Jahrbuch, pt. 1, 1899, pp. 47–62.

[2] G. Linck, Die Bildung der Oölite und Rogensteine, Neues Jahrb. Beil. B. 16, 1903, p. 495.

O. Reis, Geognostische Jahreshefte, vol. 22, 1909, p. 58.

[3] S. Wienogradski, Ueber Eisenbakterien, Botan. Zeit., vol. 46, 1888, p. 261.

E. C. Harder, Iron depositing bacteria and their geologic relations, *Prof. Pap.* 113, U. S. Geol. Survey, 1919. A recent paper containing many new and valuable data.

F. Lafar, Technical mycology, vol. 1, 1910, p. 272.

vailed; the sulphide has probably precipitated as a colloid and is, therefore, neither pyrite nor marcasite.[1]

Any of these oölitic deposits may, of course, have been enriched, after uplift and erosion, by solution of calcite but the iron was certainly not introduced by atmospheric waters.

It is very significant that the oölitic concretions in all these ores yield a delicate concentric skeleton of soft silica, upon treatment with dilute hydrochloric acid. This may indicate that a silicate had been present in all cases, but a possible alternative is that gelatinous silica and iron ore were precipitated together.

FIG. 94.—Clinton ore, Wolcott, Wayne County, New York. Magnified 20 diameters. Ore essentially formed of remains of bryozoans and crinoids. *a*, Fragments of bryozoans, calcareous walls preserved, interstices filed with ferric oxide; *b*, fragment of bryozoan encrusted with ferric oxide, the walls partially replaced by ferric oxide; *c*, bryozoan structure almost obliterated by ferric oxide; *d*, crinoid stalk replaced by ferric oxide, cells filled with calcite of uniform optical orientation; *e*, same, almost entirely replaced; *f*, calcite cement. *After L. Cayeux.*

So we arrive at the conception of shallow bays in which coral reefs flourished or the detritus of older fossiliferous limestone

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[1] Bruno Doss, Melnikovit, ein neues Eisenbisulfid, *Zeitschr. prakt. Geol.* vol. 20, 1912, pp. 453–467.

was spread. Into these bays were swept, at intervals, masses of finely divided detritus from the deep mantle of decayed rock of adjacent tropical land areas, undoubtedly rich in hematite as such products always are. The water discharged from the land certainly contained ferrous bicarbonate. In this mud agitated by the waves progressed numerous and complicated reactions. Oölites and shells of calcite were replaced by siderite, which almost simultaneously oxidized to hematite. In the deeper water glauconite was probably deposited, and it also may soon have been altered to hematite. Somewhat similar conditions are found to-day, for instance, on the south side of Molokai, Hawaiian Islands, where such hematite mud is spread out over a large area of shallow coral reef.

SEDIMENTARY MANGANESE ORES

There is much less manganese than iron in the earth's crust, the average of analyses of igneous rock calculated by Clarke showing but 0.078 per cent. of manganese. Deposits of manganese ore are also much less common than those of iron ore. Nevertheless, many spring waters carry manganese and a minute amount of it is contained in sea water. Sedimentary deposits of manganese are known, marine and lacustrine as well as fluviatile.

According to experiments by E. C. Sullivan[1] the manganese in rocks is taken into solution more easily than iron, both by carbonated water and by dilute sulphuric acid. He also finds that from mixed ferrous and manganese sulphates almost all of the iron is precipitated first by carbonate of calcium before any manganese is thrown down. Fresenius, many years ago, also found that from spring water iron is precipitated first as limonite, while the manganese remains in solution much longer. This accounts for the very general separation of the two metals in the oxidized zone.

Manganese is dissolved mainly as bicarbonate and sulphate, possibly also as phosphate. It is easily precipitated by oxidation, generally as MnO_2 in the form of pyrolusite (63.2 per cent. Mn), or as slightly hydrous psilomelane or wad (an impure mixture of manganese oxides), or more rarely as manganite ($Mn_2O_3.H_2O$). The precipitate is generally a "gel," which

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[1] E. C. Sullivan, quoted by W. H. Emmons in *Bull.* 46, Am. Inst. Min. Eng., 1910, p. 803.

crystallizes in time, but which appears to have a tendency to adsorb certain oxides, especially those of barium and potassium. According to F. P. Dunnington[1] an acid solution of ferrous sulphate dissolves manganese from the carbonate, as sulphate, with the separation of ferric sulphate and limonite; from the compound solution calcium carbonate precipitates the iron, but the manganese is precipitated only upon access of air.

$$MnSO_4 + CaCO_3 + O = CaSO_4 + MnO_2 + CO_2.$$

Bog Manganese Ore.—It has been stated above that many bog iron ores contain manganese; pure bog manganese ores are also known, though the deposits are not abundant. The material is generally earthy and soft, approaching wad in composition. In part the bog manganese consists of a skeleton of hard and glossy black ore containing cavities filled with a black powder. The deposits are rarely more than a few feet in thickness; a small occurrence near Wickes, Montana, described by Harder,[2] lies in the flat bottom of a gulch covered by soil and underlain by ochery bog limonite.

A much larger and thicker deposit occurs at Hillsborough, New Brunswick; it is said to extend over 17 acres with a thickness of 6½ feet. An analysis shows Mn, 45.81; Fe, 9.95; S, 0.03; P, 0.05, and SiO_2, 5.36 per cent.[3]

J. H. L. Vogt describes a deposit in Norway, about 1 meter thick, in a little valley above a layer of sand and below a cover of peat. The manganese ore alternates with iron ocher; it contains MnO_2, 71.20; MnO, 8.08; Fe, 1.90; P_2O_5, 0.10, and S, 0.07 per cent. In many of these occurrences the rock from which the metal was leached is a granite or a quartz porphyry.

Manganese in Lacustrine and Marine Beds.—Many sedimentary beds in all parts of the world contain manganese derived from the degradation of old land areas; it occurs as carbonate and stains or concretions of dioxide in tuffs, quartzites, sandstones, clays, shales, and limestones. It is frequently contained in beds of jasper or radiolarian chert. Strongly manganiferous sedi-

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[1] F. P. Dunnington, *Am. Jour. Sci.*, 3d ser., vol. 36, 1888, p. 177.

[2] E. C. Harder, Manganese deposits of the United States, *Bull.* 427, U. S. Geol. Survey, 1910, p. 137.

[3] *Ann. Rept.*, Geol. Survey Canada, vol. 2, 1894, p. 146.

E. C. Harder, *op. cit.*, p. 171.

ments recrystallize to crystalline schists, the manganese assuming the form of rhodonite, rhodochrosite, or manganese garnet (spessartite). The presence of manganese nodules in deep sea deposits is well known; they are considered to be rather a submarine product of segregation from the red pelagic mud than of chemical precipitation from the ocean. Very rarely, however, do these sedimentary rocks contain manganese of economic importance, and it is only by subsequent concentration, especially effective in regions of deep secular decay, that valuable deposits are developed (pp. 338–345).

An excellent example of an undoubtedly sedimentary and practically unaltered deposit is described from Newfoundland by N. C. Dale.[1] It is of little economic importance. The metal occurs as carbonate, with some MnO_2, in nodular form, in shaly and calcareous beds of Cambrian age and is associated with calcium phosphate in nodular form, hematite spherules, and barite in crystals and blades; the psilomelane in the deposit also contains baryum. Such deposits could probably only form in shallow water mud near land areas subjected to secular rock decay.

The great manganese deposits of the province of Kutais, in Trans-Caucasia,[2] are apparently sedimentary, if judged from descriptions, but it is not impossible that here, too, enrichment by decomposition has taken place. These deposits, said to be the largest in the world, are beds in Eocene clays, marls, and sandstones, the last resting on Cretaceous limestone, on the top of an extensive plateau. The ore beds, at the base of the Eocene, are 7 to 16 feet thick, and consist of several strata of oölitic pyrolusite with cementing earthy manganese ore. They are said to extend over an area of 22 square miles. The ores average 40 to 50 per cent. Mn and 0.16 per cent. P. Drake gives a complete analysis of an ore containing, MnO_2, 86.25; Mn_3O_4, 0.47; Fe_2O_3, 0.61; NiO, 0.3 per cent., and a trace of copper. Barium is present as usual in these ores. The annual production before the war was about 1,000,000 metric tons.

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[1] N. C. Dale, The Cambrian manganese deposits of Conception and Trinity Bays, Newfoundland, *Proc.*, Am. Philos. Soc., vol. 54, 1915, pp. 371–456.

[2] C. F. Drake, The manganese ore industry of Caucasus, *Trans.*, Am. Inst. Min. Eng., vol. 28, 1898, p. 191.

E. C. Harder, *op. cit.*, p. 208.

SEDIMENTARY PHOSPHATE BEDS[1]

Composition of the Calcium Phosphates.—Phosphorus enters in the average composition of igneous rocks, according to F. W. Clarke, to the extent of only 0.11 per cent., and the analyses of sediments show smaller percentages. Nevertheless, it plays a most important part in the life processes of plants and animals, in the sea and on the land, and in places its compounds accumulate in large masses. Its most common salt is a calcium phosphate; the phosphates of iron, aluminum, lead, and other metals are entirely subordinate.

Apatite, the most common calcium phosphate, also contains CaF_2 or $CaCl_2$. The formulas may be written $Ca_5(PO_4)_3F$ and $Ca_5(PO_4)_3Cl$, or $3Ca_3(PO_4)_2.Ca(F,Cl)_2$, the first part of the latter formula being the tri-basic calcium phosphate. Fluorine apatite contains 42.3 per cent. P_2O_5; chlorine apatite, 41.0 per cent. The pure tri-basic phosphate, which is used as a standard to express the tenor of phosphate rocks, contains 45.8 per cent. P_2O_5. The phosphate in sedimentary rocks approaches more or less closely the tri-basic phosphate, but sometimes is almost identical with a fluorine apatite.

In deposits of guano a considerable number of acid hydrous phosphates such as, monetite ($CaH.PO_4$) and brushite ($CaH.PO_4.2H_2O$) have been found,[2] but they have little practical importance. In the same deposits various complex phosphates of

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[1] R. A. F. Penrose, Jr., Nature and origin of deposits of phosphate of lime, *Bull.* 46, U. S. Geol. Survey, 1888. (Gives bibliography.)

David Levat, Étude sur l'industrie des phosphates, *Ann. des Mines*, 7, 1895, 135.

X. Stainer, Bibliographie générale des gisements de phosphates, *Ann. des Mines de Belgique*, vol. 7, 1902, *et seq.*

F. W. Clarke, The data of geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, pp. 519–528.

O. Stutzer, Die wichtigsten Lagerstätten der Nicht-Erze, Berlin, 1911, pp. 265–461.

Eliot Blackwelder, The geologic rôle of phosphorus, *Am. Jour. Sci.*, 4th ser., vol. 42, 1916, pp. 285–298.

W. C. Phalen, The conservation of phosphate rock in the United States, *Trans.*, Am. Inst. Min. Eng., vol. 57, 1918, pp. 99–132.

Mineral Resources, U. S. Geol. Survey, Annual publication; various authors.

Mineral Industry, New York, Annual publication, various authors.

[2] F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 520.

iron, magnesium, sodium, and ammonium occur, but these also are unimportant.

The mineralogical composition of the marine and residual phosphates is complex.[1] Apatite is essentially a high temperature mineral and has not been recognized in the marine phosphates; in the latter hydrous carbono-phosphates play the principal part. The latter are amorphous and doubtless hardened colloid precipitates; they are referred to two species: collophanite ($9CaO.3P_2O_5CaO.CO_2.H_2O + nH_2O$) and fluocollophanite. The crystalline minerals which in part are altered colloids, in part crusts and mammillary structures comprise dahllite and francolite both of which are similar carbono-phosphates with or without fluorine.

The marine phosphate rocks, aside from detrital impurities, contain thus calcium carbonate and calcium phosphate; shell fragments and glauconitic granules are frequently present. The poorer kinds may be classified as phosphatic sands, marls, or limestones. The richer varieties are usually oölitic, dark-colored rocks, occasionally with a peculiar whitish efflorescence, and may carry large amounts of organic matter. They are inconspicuous and in places difficult to recognize. The specific gravity, averaging 2.9 in 70 per cent. phosphate rock, is considerably higher than that of limestone and may be used to aid in the identification. A rapid field assay with ammonium molybdate is the best test.

Other Phosphates.—Among the iron phosphates, vivianite, $Fe_3(PO_4)_2.8H_2O$, is the best known, and it appears frequently in bog iron ores. Of the aluminum phosphates, wavellite, $4AlPO_4.2Al(OH)_3 + 9H_2O$, and turquoise, $AlPO_4.Al(OH)_3 + H_2O$, are the best known, the former locally used as a source of phosphorus, the latter a blue semi-precious stone; both are usually products of the uppermost zone of the crust, sometimes even forming in the zone of oxidation. In a similar geological position occur the lead phosphate, pyromorphite, corresponding in formula

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[1] A. Lacroix, Sur la constitution mineralogique des phosphorites francaises. Compte Rendu, vol. 150, 1910, p. 1213.

H. S. Gale and R. W. Richards, *Bull.* 430, U. S. Geol. Survey, 1910, p. 464.

W. T. Schaller, *Bull.* 509, U. S. Geol. Survey, 1912, pp. 89–100.

A. F. Rogers, *Am. Jour. Sci.*, 4th ser., vol. 33, 1912, p. 475.

A. F. Rogers, A review of the amorphous minerals, *Jour. Geol.*, vol. 25, 1917, pp. 515–541.

to chlorine apatite. Other phosphates, like amblygonite, a fluo-phosphate of lithium and aluminum, monazite, and other phosphates of the rare earths, find their home in the pegmatite dikes. This illustrates the variety of occurrence of the phosphates.

Phosphate Deposits.—The many kinds of deposits in which calcium phosphate is of economic importance are shown by the following list:

1. Disseminated in igneous rocks or in their differentiation products of metallic ores.
2. Apatite veins, closely allied to pegmatitic dikes.
3. Marine concretionary beds.
4. Sub-aërial accumulations of animal excrements—bat caves, guano islands.
5. Metasomatic deposits by replacement of limestone by means of phosphate solutions, from Nos. 3 and 4.
6. Residual concretions, by action of atmospheric waters on No. 3.

Use.—The principal use of the calcium phosphate is for soil fertilization, and all the classes enumerated above are so utilized. Under No. 1 comes, for instance, the apatite concentrate from the Adirondack magnetite ores; under No. 2 the apatite veins of Canada and Norway; the occurrences of the remaining classes are described below.

For utilization it is necessary to transform the insoluble tri-basic phosphate into soluble form and this is generally effected by a 60 per cent. solution of sulphuric acid;[1] hence the dependence of the phosphate industry on an abundant and cheap supply of sulphuric acid, illustrated, for instance, in the establishment of large sulphuric acid plants at the pyritic copper deposits of Ducktown, Tennessee, for the treatment of the sedimentary phosphates of the Southern States. The treatment with H_2SO_4 results in a partial decomposition, with the formation of soluble calcium phosphate, also called super-phosphate or mono-calcium phosphate ($Ca.H_4(PO_4)_2.H_2O$); and also some di-calcium phosphate, which is much less soluble. The standard is 77 per cent. of the tri-basic calcium phosphate with less than 3 per cent. of alumina plus iron, but not all of the production reaches this grade.

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[1] The reaction is expressed by the following formula: $Ca_3(PO_4)+2H_2SO_4 = CaH_4(PO_4)_2+2CaSO_4$.

Experiments show that even the tri-calcium phosphate or apatite is soluble, particularly in water containing carbon dioxide; its solubility in solutions of $CaCO_3$ or in pure water is slight, but the presence of sodium chloride increases the solubility.[1] The marked absorption of phosphoric acid by clays and soils is held to be due to the presence of colloid bodies.

Production.[2]—Though some phosphates are obtained from apatite deposits and from basic slags, the greater part comes from sedimentary and residual beds. In the United States, the bulk of the production comes from Florida, Tennessee, and South Carolina, in the order named; by far the most is mined in Florida. The yield of the United States in 1913 was about 3,000,000 long tons. Large quantities were exported. In 1917 owing to war conditions the output was one-third less. Scarcity of sulphuric acid and reduced exports contributed to this. The average price of Florida phosphate was $3 per ton in 1916.

Of other countries Algeria and Tunis produced about 2,800,000 metric tons, and France about 400,000 tons in 1913, but these figures have been greatly reduced since the war began. The production of the guano islands of the Pacific is now comparatively unimportant.

Origin of the Phosphate Rocks.[3]—As all land animals absorb phosphoric acid and segregate it as calcium phosphate in their bones and excrements, it is not difficult to understand the accumulation of phosphates wherever animal life is particularly abundant and undisturbed. Besides phosphates, such deposits contain much ammonia and nitrogen, except where subjected to leaching by heavy precipitation. Of this kind are the bone beds which are found occasionally in various formations and in caves.

The guano of commerce is deposited by sea birds congregating in enormous numbers on desert coasts and oceanic islands, for instance, along the Peruvian and Chilean coast, on Christmas Island in the Indian Ocean, and in the West Indies. Some

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[1] H. E. Patten and W. H. Waggaman, Absorption by soils, *Bull.* 52, Bureau of Soils, Dept. Agriculture, 1908.

O. Schreiner and G. H. Failyer, Absorption of phosphates and potassium by soils, *Bull.* 32, Bureau of Soils, Dept. Agriculture, 1906.

[2] *Mineral Resources*, U. S. Geol. Survey, annual issues.

[3] Sometimes described as "phosphorites" (Stelzner and Bergeat, 1, p. 442). The name phosphate rock seems more appropriate, especially as some authors (Merrill, Non-metallic minerals, 1910, p. 267) use phosphorite in a somewhat different sense.

of these deposits cover whole islands and in places may accumulate to a depth of 100 feet, and it is stated that under favorable circumstances the rate of deposition is rapid. The guano of dry climates varies greatly in texture and color, but generally is granular, light colored, and porous. It contains on an average 10.90 per cent. nitrogen, 27.60 per cent. phosphates, and 2 to 3 per cent. potash.[1]

The West Indian deposits—for instance, those on Navassa[2] and Sombrero islands—have been leached and are in part hard and compact, in part porous and friable. The phosphate has been concentrated to 70 or 75 per cent. The material contains from 21 to 40 per cent. of phosphoric acid, 1 to 2 per cent. sulphuric acid, 20 to 45 per cent. lime, usually also much ferric oxide and alumina. The underlying limestone or igneous rock may be locally replaced by the phosphatic solutions.

The marine phosphate beds also derive their material from animal life. Sea water contains phosphoric acid, though the quantity is extremely small, and likewise some fluorine, each amounting to about a little less than one part per million.[3]

According to Carnot, many shells, particularly those of the older formations, are rich in phosphorus and fluorine. A Cambrian *Obolus* contained 36.54 per cent. P_2O_5 and 2.78 per cent. F; a recent *Lingula* yielded 23.20 per cent. P_2O_5 and 1.52 per cent. F.[4] The shells of crustaceans[5] contain up to 26 per cent. $Ca_3P_2O_8$. Pteropods, lamellibranchs, gastropods and protozoans also carry phosphorus. Corals likewise contain a small amount of phosphorus and fluorine, and the same substances are found in the bones and teeth of fishes. The marine sediments, then, all hold more or less of phosphates, and it is a matter of some surprise that fluorite does not more commonly occur in sedimentary rocks.[6]

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[1] R. A. F. Penrose, Jr., *Bull.* 46, U. S. Geol. Survey, 1888.

[2] E. V. D'Invilliers, Phosphate deposits of the Island of Navassa, *Bull.*, Geol. Soc. Am., vol. 2, 1891, p. 71.

[3] A. Carnot, *Ann. des Mines*, 9th ser., vol. 10, 1896, p. 175.

[4] Andersson and Sahlbom, Ueber den Fluorgehalt schwedischer Phosphorite, *Bull.* 4, Geol. Inst. Upsala, 1900, p. 79. Neues Jahrb., ref., 1903, 1, pp. 195, 197.

[5] F. W. Clarke and G. Steiger, *Proc.* Nat. Acad. Sci., vol. 5, 1919, pp. 6–8.

[6] K. Andrée, Ueber einige Vorkommen von Flusspath in Sedimenten, *Tsch. M. und p. Mitt.*, vol. 28, 1909, pp. 535–562.

H. S. Gale and R. W. Richards, *Bull.* 430, U. S. Geol. Survey, 1910, p. 463.

In some beds the phosphates occur disseminated in small quantities, in part as small concretions, in part remaining in the shell fragments. In the more valuable deposits the phosphates appear in more concentrated form and characteristically assume the forms of nodules, or concretions (sometimes of large size), or oölitic rocks built up of small oölites in part of concentric and fibrous structure. The nodules have often a shell nucleus and, as a result of enrichment, may contain more phosphate in the peripheral than in the central parts.

While phosphate nodules have been brought up by the dredge from great oceanic depths, the conditions for their formation are probably best at moderate depths, near shores, where the marine life is most abundantly developed or, as pointed out by some authors, where sudden changes of temperature, owing to conflicting currents, kill large numbers of marine organisms.

The origin of the oölitic and nodular phosphate rocks, in some of which recognizable organic remains are scarce, has been discussed extensively, but is as yet not fully explained. It is believed that ammonium phosphate may form in the organic matter and that this reacts on shell remains, replacing them with calcium phosphate, which eventually accumulates in larger concretions.[1] These processes are likely to continue for some time at least after the sedimentation, in the yet soft sediments.[2]

After the beds have been uplifted and exposed to weathering enrichment takes place easily, by the removal of calcium carbonate. This is especially effective in regions of deep rock decay, as in the Southern States. The rock phosphates of Utah and Idaho have remained almost unaltered.

The cycle of migration of the phosphates is a fascinating study. From their original home in the igneous rocks they are dissolved by surface waters and absorbed by all living things, vegetable and animal, on land and on sea. After the death of the organisms the phosphates return to the soil or to the sedimentary beds to be dissolved and used anew by other generations.

[1] Renard and Cornet, *Bull.* 21, ser. 3, Acad. Belgique, 1891, p. 126.

L. Kruft, *Neues Jahrbuch*, Beil. Bd. 15, 1902, pp. 1–65, Ref. in *Zeitschr. prakt. Geol.*, vol. 10, 1902, p. 301.

R. Delkeskamp, *Zeitschr. prakt. Geol.*, vol. 12, 1904, p. 299.

[2] L. Cayeux, for instance, presents a figure, showing a small concretion of phosphate molded against a grain of glauconite; the latter itself being formed after the sedimentation; Contrib. à l'étude micrographique des terrains sédimentaires, *Mém.* Soc. géol. du Nord, 4, pt. 2, Lille, 1897.

Occurrences of Phosphate Rocks.—Deposits of phosphate rock are found in the marine beds of all ages and in almost all countries, at least from the Cambrian, when the segregation of phosphoric acid by the inhabitants of the sea appears to have begun, to the Tertiary, and in the present oceans such deposits certainly continue to form. In description it is impracticable to separate the primary marine deposits from those altered by weathering.

Large deposits, enriched by weathering, are worked in the Cretaceous beds of northern France. In the southwestern part of that country, in the departments of Lot and Lot-et-Garonne, phosphates occur in irregular fissures with clay in Jurassic limestone.[1] These deposits are probably formed by replacement effected by descending solutions from sedimentary phosphate beds.

Phosphate beds are now mined on a large scale along the frontier of Algeria and Tunis.[2] The beds occur in the lower Eocene, which covers Cretaceous strata, and consist in part of larger concretions in marl, sometimes carrying the rich phosphate only as a crust; other beds are formed by a soft material, consisting of small and smooth brown or yellowish grains of phosphate cemented by calcite and also containing many fossils and much bituminous matter. The thickness of the richest phosphatic stratum is said to be 10 to 15 feet.

The deposits found in the United States are mainly in three regions—(1) the Atlantic coast belt of Tertiary rocks in the Carolinas and Florida; (2) the Tennessee area of Silurian and Devonian strata; (3) the Utah-Idaho region of Carboniferous beds.

The phosphates of the Utah-Idaho region[3] were discovered only recently, but are of great extent and prospective value; at

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[1] L. de Launay, Gites mineraux, vol. 1, 1913, p. 679.

[2] M. Blayac, Description géologique de la région des phosphates du Dyr et du Kouif, *Ann. des Mines* (9), 6, 1894, pp. 319–330.

L. de Launay, Les richesses minerales de l'Afrique, Paris, 1903, p. 206.

O. Tietze, Die Phosphatlagerstätten von Algier und Tunis, *Zeitschr. prakt. Geol.*, 1907, p. 229.

[3] F. B. Weeks and W. F. Ferrier, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 449–462.

H. S. Gale and R. W. Richards, *Bull.* 430, *idem,* 1910, pp. 457–535.

Eliot Blackwelder, *Bull.* 430, *idem,* 1910, pp. 536–551.

R. W. Richards and G. R. Mansfield, *Bull.* 470, *idem,* 1911, pp. 371–439.

G. R. Mansfield, *Am. Jour. Sci.*, 4th ser., vol. 46, 1918, pp. 591–598.

present, owing to difficulties and cost of transportation, they are mined only on a small scale near Montpelier, Idaho.

They extend north of Ogden, Utah, into Idaho, Wyoming and Montana, and the best deposits are in the ranges which constitute the northern continuation of the Wasatch. Their position is in the Park City formation of the Upper Carboniferous (Pennsylvanian), which has an average thickness of 600 feet and consists of limestones, cherty in part, phosphate beds, and shales. The phosphate horizon is in the middle of the formation and the beds have an average thickness of 200 feet. (See Figs. 8, 9, and 95.) The rocks are massive brown to gray phosphatic shales and beds of rock phosphate with some limestone. The richest bed mined at Montpelier, carrying 70 per cent. or more of $Ca_3P_2O_8$, lies at the base of the phosphate section and is 5 or 6 feet thick. It is a black to dull-gray oölitic rock, with grains of all sizes up to pebble-like bodies one-half inch in diameter.

Large sections of the phosphatic beds, in places a thickness of 75 feet, carry from 30 to 50 per cent. of $Ca_3P_2O_8$. The beds are folded and locally have steep dips. The rock is hard and the mining is carried on by underground operations. Very little enrichment is noted.

The phosphates of western Tennessee[1] have been worked since 1894 and at present yield about 400,000 tons per annum. They are of three classes. 1. Brown residual phosphates, resulting from leaching of Ordovician phosphatic limestones; the beds are from 3 to 8 feet thick and carry as much as 80 per cent. of tri-calcium phosphate. 2. The blue or black bedded phosphates, which occur in beds of Devonian age, show variations from oölitic through compact and conglomeratic to shaly forms. The high-grade rock is seldom more than 20 inches thick. The nodular variety, which is embedded in a green sand, carries about 60 per cent. $Ca_3P_2O_8$. 3. The white phosphate, which is a post-Tertiary product of replacement or filling of cavities of limestone of Carboniferous age. None of it is now mined.

The phosphate beds of North and South Carolina,[2] discovered

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[1] C. W. Hayes, The Tennessee phosphates, *Sixteenth Ann. Rept.*, U. S. Geol. Survey, pt. 4, 1895, pp. 610–630; *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, pp. 513–550; *Twenty-first Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1901, p. 473.

[2] G. S. Rogers, Phosphate deposits of South Carolina, *Bull.* 580, U. S. Geol. Survey, 1914, pp. 183–220.

in 1867, extend along the coast for a distance of 60 miles. They are contained in loose beds of Miocene age, rich in fossils. The land deposits lie at a shallow depth and consist of so-called pebble rock, a solid mass from which the calcium carbonate has been leached and partly replaced by phosphate; the solution cavities give this material the appearance of a mass of separate pebbles.

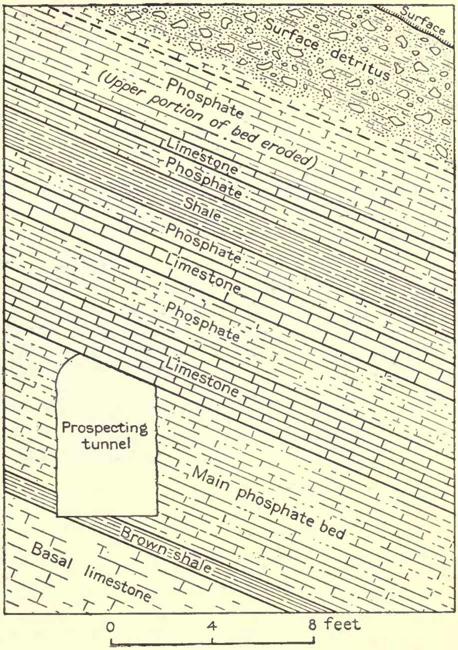

FIG. 95.—Section showing beds of phosphate, Montpelier, Idaho. *After Weeks and Ferrier, U. S. Geol. Survey.*

The rock varies from 1 to 3 feet in thickness and is covered by a green sandy marl. Similar deposits have been dredged in the rivers; they consist essentially of water-rounded fragments of the land rock. The mining is carried on by steam-shovel or dredge operations.

The phosphate deposits of Florida[1] are at present the most productive in the world and large quantities are, under normal conditions, exported to almost all European countries.

The deposits follow in the main the northwestern coast of the State but lie some distance from the shore. There are several types, all contained in the Alum Bluff formation or above it, or in the underlying Vicksburg limestone, both formations of Oligocene age. The clays, marls and sandstones of the Alum Bluff contain in several horizons abundant smooth yellowish or brown nodules or ovules of phosphate which are considered by Matson and Sellards as of primary deposition and the source of all the other deposits. These beds are worked in some places but are not of great importance. The material is very similar to the phosphates of Gafsa in Tunis.

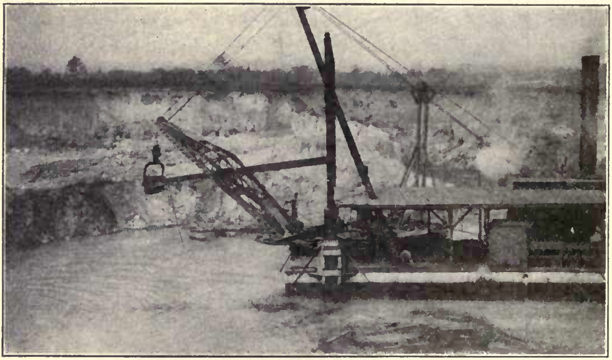

FIG. 96.—Dredging Florida phosphates. Upper bench is sandy overburden. *Photograph by F. B. Van Horn.*

The so-called "land pebble" deposits which are the most important are rudely stratified detrital and residual masses; they rest on the Alum Bluff formation and are believed to be derived from this source. They are believed to be of Miocene or

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[1] E. H. Sellards, *Fifth Ann. Rept.*, Florida Geol. Survey, 1913, pp. 23–80.

E. H. Sellards, *Trans.*, Am. Inst. Min. Eng., vol. 50, 1914, pp. 901–916.

G. C. Matson, The phosphates of Florida, *Bull.* 604, U. S. Geol. Survey, 1915.

See also successive issues of *Mineral Resources*, U. S. Geol. Survey, and *Mineral Industry.*

Pliocene age. They contain concretions of white phosphate, averaging 65 to 70 per cent. tribasic calcium phosphate, while the finer matrix often contains 20 to 30 per cent. of the same. They lie at elevations of about 100 feet and form parts of gravel beds with casts of shells, shark's teeth, and bones of mastodon, horse and rhinoceros. The average depth of deposit is about 12 feet, that of the sandy overburden up to 40 feet. The workable deposits average several acres, though some cover as much as 40 acres.

The type of deposit called "rock phosphate" occurs in the Vicksburg limestone of lower Oligocene age and rest in depressions on its surface. They are of secondary origin and are believed to be leached from the overlying Alum Bluff formation. The concentration was affected by chemical and mechanical means; the result is a mass of rounded or subangular phosphate concretions ("rocks" or "pebbles") in places rudely sorted in layers of coarse and fine, and contained in a matrix of sand, clay or soft phosphate.

Pleistocene sand, up to 50 feet thick, covers the deposits. There are many deposits covering from 5 to 35 acres, while some are mere pockets in the limestone. The average thickness is 30 feet; transitions to the underlying limestone are sometimes observed. The concretions or nodules vary from a few inches to 10 feet in diameter, are close grained and light gray and sometimes show cavities lined with secondary and mammillary phosphate. Shells are rare but fragments of bones and shark's teeth occur here and there. The processes which have operated in the concentrations from the overlying Alum Bluff beds are replacement of limestone by phosphate and solution of residual limestone with resulting slumping; in places there has been mechanical transportation. The soft phosphates are thought to be formed by replacement of porous limestone. The percentage of "recoverable" phosphate in the deposits does not average much above 15 per cent.

The so-called "river pebbles" are Pleistocene deposits in the present rivers but are not worked now.

Mineralogically, the Florida phosphates are held by Matson to consist of collophanite with francolite (p. 284).

The material is mined by steam shovel, by hydraulic method or by dredge (Fig. 96); it is then washed to remove the clay and afterward crushed, screened and hand-picked.

The chemical composition of the marine and residual phosphates is shown in the following analyses of which I and III represent unaltered marine deposits and II a residual occurrence.

ANALYSES OF PHOSPHATES

	I	II	III
Insoluble	1.82	6.69	3.05
SiO_2	0.30	.....	......
Al_2O_3	0.50	2.14	1.09
Fe_2O_3	0.26	0.61	0.64
MgO	0.22	0.33	0.57
CaO	50.97	46.03	48.58
Na_2O	2.00	......	0.09
K_2O	0.47	......	
H_2O-	0.48	0.79	......
H_2O+	0.57	3.47	......
CO_2	1.72	3.93	4.60
P_2O_5	36.35	31.50	29.74
SO_3	2.98	......	2.75
Cl		......	0.11
Fl	0.40	1.86	2.12
Organic		......	7.45
	99.04	97.35	100.79

I. Crawford Mountains, Utah. Geo. Steiger, analyst, *Bull.* 430, U. S. Geol. Survey, 1909, p. 465. No titanium, organic matter not determined, trace chlorine.

II. Florida. Land pebble, G. H. Eldridge.

III. Gafsa, Tunis. O. Tietze, *Zeitschr. prakt. Geol.*, 1907, p. 248.

Analysis calculated on dry material, 3.81 per cent. H_2O. P_2O_5 equivalent to 64.93 per cent. $Ca_3P_2O_8$; 2.35 per cent. $CaFl_2$; 4.67 per cent. $CaSO_4$; 10.45 per cent. $CaCO_3$.

CHAPTER XVII

DEPOSITS FORMED BY EVAPORATION OF BODIES OF SURFACE WATERS

THE SALINE RESIDUES[1]

INTRODUCTION

The deposits thus far described have been in the nature of insoluble residues, or chemical precipitates of relatively insoluble substances in lakes, rivers, and seas. There are, however, other deposits which also may be considered as chemical precipitates in surface waters but which consist of soluble salts formed by the evaporation of waters in closed or partially closed basins. They contain the easily soluble substances leached from the crust, brought down by the rivers to oceans and lakes, and finally concentrated under certain characteristic conditions.

Closed basins are typical of dry climate and of deserts. The slow crustal movements tend to create them everywhere, by folding, subsidence, and uplift, but in the deserts the streams have not the power to cut outlets and to keep the drainage lines established. On the contrary, the movement of the débris from the mountain ranges in broad alluvial fans or aprons increases the tendency toward closed basins. The dry climate accelerates evaporation and the precipitation of the salts; dust storms transport vast masses of fine detritus; blinding salt flats extend between the barren mountain chains. Thus, at present, salt beds are found in the Cordilleran deserts along the western side of the whole American continent, in the Sahara, and in the arid, central part of Asia. Similar conditions existed in the past in different parts of the world: The Permian in central Europe, the Triassic in the Rocky Mountain region, and the Silurian in eastern North America—all these ages were at times characterized by arid wastes and deposition of salt and gypsum.

Saline deposits may then form: (1) in bays of the sea; (2) in lakes;

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[1] George P. Merrill, The non-metallic minerals, 1910.

F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, pp. 217-259.

(3) in playas or intermittent lakes; (4) on arid slopes by rapid evaporation of storm waters.

In the latter two classes capillary ascent of the solutions often help to bring the salts to the surface to form "efflorescences" or saline crusts as illustrated by the crusts of "alkali" (carbonate, chloride and sulphate of sodium, sulphate of calcium) which so often interfere with agriculture in dry countries.

In regions of calcareous rocks, as in the undrained basins of Mexico, soft or compact beds of calcium carbonate locally called "caliche" or "tepetate" often cover the gentle slopes below the mountains. These belong in class 4. Minor saline deposits may result from evaporation at the surface of waters from ascending springs.

Bodies of soluble salts are rarely formed below the surface; but reactions may take place in a buried deposit by which new salts are formed or concentrations of disseminated substances are effected. In places it may be difficult to distinguish these strictly speaking epigenetic bodies from the syngenetic salts (p. 313).

No traces of metallic ores are found in the saline residues. Minute amounts of gold and silver have been found in salt from sea water (p. 13). Regarding traces of gold in the potassium deposits in Germany the evidence is conflicting.[1]

TYPES OF WATER

From a geological standpoint there are two types of water in the seas and closed basins. The first, which may be called the oceanic type, contains dominant sodium chloride and is characteristic of the sea as well as of partly evaporated lakes in regions where sedimentary rocks prevail; the Great Salt Lake of Utah is an example. When such water is subjected to extreme evaporation, as in the Dead Sea, a "residual" type rich in magnesium chloride results. The second main type is that of generally smaller closed basins in regions of great volcanic activity; this type contains an abundance of sulphate and carbonate of sodium, besides more or less chloride; it indicates the result of the first leaching of loose volcanic ejecta and also shows the influence of the discharge of hot springs containing

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[1] A. Liversidge, *Jour.* Chem. Soc., vol. 71, 1897, p. 298.
E. E. Lungwitz, *Eng. and Min. Jour.*, April 6, 1905.

sodium carbonate and borate. The water of Mono Lake, California, is a good example.

Certain sedimentary series, such as the Cretaceous of the Western States, contain abundant alkaline sulphates. Leaching of these beds by atmospheric waters takes place and these products may be carried down into salt flats and small lakes. By chemical reactions (p. 59) carbonate of sodium forms from other sodium and calcium salts and the lakes often contain much of this salt besides the sulphates. Such alkali lakes occur in Wyoming, for example. Borates characteristic of volcanic regions are generally lacking in these lakes.

The first, "oceanic" type of waters yields deposits of gypsum, common salt, and finally potassium and magnesium salts. The second, "volcanic" type yields soda, glauber salt, borates, probably also nitrates, as well as more or less sodium chloride.

COMPOSITION OF SALTS IN WATER OF SEA AND CLOSED BASINS

	I	II	III	IV	V
Cl	55.292	55.69	70.25	10.45	23.34
Br	0.188	tr.	1.55	....	....
SO_4	7.692	6.52	0.21	54.07	12.86
B_4O_7	......	.....	.....	.....	0.32
CO_3	0.207	.....	tr.	4.24	23.42
Li	......	0.01	.....	.....	.....
Na	30.593	32.92	6.33	25.88	37.93
K	1.106	1.70	1.70	......	1.85
Ca	1.197	1.05	5.54	tr.	0.04
Mg	3.725	2.10	14.42	5.36	0.10
Fe_2O_3	......	.....	tr.	......	tr.
Al_2O_3	......	.....	tr.	......	tr.
SiO_2	......	0.01	tr.	......	0.14
	100	100	100	100	100

I. Oceanic type. Average of 77 analyses, W. Dittmar, Challenger Rept., vol. 1, 1884. Salinity 3.5 per cent.

II. Oceanic type. Great Salt Lake, E. Waller, School of Mines, Quart., vol. 14, 1892. Salinity 23 per cent., subject to variations.

III. Residual type. The Dead Sea, Compt. Rend., vol. 62, 1866, p. 1329. A. Tereil, analyst. Salinity 20.7 per cent.

IV. Sulphate type. Devil's Lake, North Dakota, F. W. Clarke, Geochemistry, 1916, p. 163; H. W. Daudt, Analyst. Salinity 1.1 per cent.

V. Volcanic type. Mono Lake, *Bull.* 60, U. S. Geol. Survey, 1890, p. 53. T. M. Chatard, analyst. Salinity 5.1 per cent.

The general relation of the salts dissolved in oceanic waters to those in rivers is as follows:

In ocean waters: $Cl > SO_4 > CO_3$; $Na > Mg > Ca$
In river water: $CO_3 > SO_4 > Cl$; $Ca > Mg > Na$.

Change from river to lake water involves a concentration of chlorides and relative loss of magnesia, silica, and lime.

NORMAL SUCCESSION OF SALTS

When water evaporates until precipitation of the dissolved salts begins, the least soluble salts will generally fall down first, while the most soluble salts will remain in the solution until the last. Experiments by J. Usiglio[1] on sea water showed that the carbonates of calcium and magnesium, with a little ferric oxide, were precipitated when one liter of the water was reduced from one-half to one-fifth. Gypsum was precipitated when the volume was one-fifth to one-seventh, but continued in lessening amounts until only 30 cubic centimeters of the original liter remained. Sodium chloride was precipitated abundantly upon reduction of volume to 100 cubic centimeters, but continued until the volume of the water was only 16 cubic centimeters; even then some of the salt remained in solution. Chloride and sulphate of magnesium fell down within the same limits but in increasing quantities, and the residual "bittern" contained mainly the chlorides of magnesium and potassium, bromide of sodium, sulphate of magnesium, and chloride of sodium. Naturally the quantity of $NaCl$ greatly exceeded that of the other salts.

The whole series of these salts is rarely represented in the saline deposits; the best known and almost only example of such complete evaporation is found in the great Prussian potash and salt deposits. Often, as in the "Red Beds" of the Western States, the process ceased after the gypsum was laid down, and changes of climate or invasion of the sea may have prevented the formation of sodium chloride.

Actually the conditions and the results of precipitation are far more complex than the experiments mentioned would seem to show. The influence of temperature and time may vary the details of the precipitation greatly, and double or complex salts are often formed.

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[1] *Annales chim. phys.*, 3d ser., vol. 27, 1849, pp. 92–172.

A saline solution containing the same salts as sea water but in different proportions would yield materially unlike results upon evaporation. In brief, temperature, concentration, and time are always factors of great importance in the origin of saline residues.

The study of the stability fields of these salts has received much impetus by the labors of J. H. van't Hoff and his numerous associates,[1] undertaken mainly to elucidate the problems of the potassium deposits of Prussia.

The occurrence of thick beds of anhydrite is explained by the work of van't Hoff and Weigert,[2] who established that the mineral forms from gypsum in sodium chloride solutions at 30° C. In sea water the transformation takes place at 25° C. Crystals of gypsum, sinking through a salt solution at that temperature, are converted into anhydrite. This is an example of the more general rule of dehydration of minerals in contact with salt solutions, at temperatures considerably below their normal inversion temperature.

Many minerals are deposited in nature from solutions in a lower state of hydration than is produced at ordinary temperatures in the laboratory. Thus natron, the ordinary sodium carbonate ($Na_2CO_3.10H_2O$) rarely occurs as a natural product, although it is stable at temperatures below 37° C. The more common product is trona ($Na_2CO_3.NaHCO_3.2H_2O$), which is ordinarily stable only above 37° C.

From pure sodium sulphate solution mirabilite ($Na_2SO_4.10H_2O$) is ordinarily deposited, but in the presence of sodium chloride thenardite (Na_2SO_4) is formed. From a solution of magnesium sulphate in the presence of magnesium chloride kieserite ($MgSO_4.H_2O$) is precipitated instead of the heptahydrate. The presence of a co-solute, by lowering the osmotic pressure, acts in the same direction as a rise of temperature.[3] This principle is undoubtedly also applicable to minerals in rocks and veins and explains many anomalies of mineral occurrence.

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[1] Mainly published in the *Sitzungsberichte* K. preuss. Akad. d. Wiss., from 1897 to the present time.

The results are summarized by van't Hoff in a book entitled "Zur Bildung der Oceanischen Salz-Ablagerungen," Braunschweig, 1905, and 1909, and in "Physical chemistry in the service of the sciences," Univ. Chicago Press, 1903.

[2] *Sitzungsber.* Akad., Berlin, 1901, p. 1140.

[3] J. V. Elsden, Principles of chemical geology, 1910, pp. 85–86

STRUCTURAL FEATURES

In desert valleys filled by temporary or permanent lakes we often find a succession of salt beds of no great thickness alternating with detrital matter of eolian or fluviatile origin. If the basin is large and deep a considerable mass of salt may accumulate. The laws governing the deposition of saline residues in the depressions of the deserts have been ably set forth by J. Walther.[1]

The salt deposits of marine origin are frequently of great thickness. In some cases they aggregate 1,000 to 2,000 feet, and it will be readily recognized that any theory based on a single cycle of evaporation of sea water, containing only 3.5 per cent. of salts would meet with great difficulties.

The "bar theory," presented by C. Ochsenius[2] in 1877, but already suggested by previously expressed views of Miller, Lyell and Bischof, attempts to explain these thick salt beds. Ochsenius believed that salt deposits of the purity and thickness of those in central Germany could not have been formed by the flooding of a series of shallow sounds and lakes. A periodical repetition of evaporation and flooding would necessitate an improbable number of uniform epochs of subsidence and elevation.

Walther's theory involving a leaching of the small saline content of old sediments and its gradual concentration in desert valleys was rejected by Ochsenius as not properly representing the conditions prevailing in the saline deposits of central Germany.

The bar theory premises a bay of the ocean separated from the open sea by a practically level bar which permits only about the same quantity of water to enter as is evaporated from the surface. A dry climate and absence of fresh-water tributaries to the bay are also premised. Under these conditions the sea water entering over the bar continuously carries a new supply to the bay; the surface layers, becoming denser, always sink and the concentration continually increases in the enclosed body of water.

The salt deposits on the bottom are increasing in thickness and the heavy "bittern" solution, with the remaining magnesium salts, correspondingly rises toward the surface.

When these dense solutions reach the surface of the bar, the

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[1] J. Walther, Lithogenesis der Gegenwart, Jena, 1893, 1894, pp. 776–800. J. Walther, Das Gesetz der Wüstenbildung, Leipzig, 1912.

[2] C. Ochsenius, Die Bildung der Steinsalzlager, Halle, 1877, p. 172. C. Ochsenius, Bedeutung des orographischen elementes "Barre," *Zeitschr. prakt. Geol.*, 1893, pp. 189–201; 217–233.

movement is reversed and the residual "bittern" will flow outward into the ocean. No accumulation of potassium-magnesium salts will occur. Should, however, the bar have increased in height just at this time, the bittern would be retained in the now closed basin and the deposition of the potassium salts would follow.[1]

The Gulf of Karaboghaz, on the eastern side of the Caspian Sea, is frequently referred to as an excellent illustration of the bar theory.[2]

GYPSUM AND ANHYDRITE[3]

Occurrence.—Gypsum ($CaSO_4.2H_2O$) and anhydrite ($CaSO_4$) usually occur in sedimentary beds as saline residues. Both are also found occasionally as gangue minerals in ore deposits and gypsum is in places a product of ascending springs or of reactions of acid waters on calcareous beds. Efflorescences of gypsum may be produced by capillary action over gypsiferous beds or along saline lakes. Anhydrite, on account of its slow transformation into the hydrous compound, has no economic value, while gypsum is one of the most important non-metallic minerals.

Gypsum in sedimentary deposits frequenty forms almost pure beds of considerable thickness. It appears as snow white fine-grained aggregates; characterized by softness (H:2), low specific gravity (2.3), perfect cleavage and great solubility in dilute hydrochloric acid. Anhydrite likewise forms white granular aggregates, but is easily distinguished from gypsum by its greater hardness, its greater specific gravity (2.9), its pseudocubical cleavage and resistance to weak HCl. Anhydrite

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[1] H. Everding, Deutschlands Kalibergbau, Berlin, 1907, pp. 37–40.

[2] F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 165.

[3] F. A. Wilder, *Eng. & Min. Jour.*, vol. 74, 1902, p. 276; *Mines and Minerals*, Dec., 1909; *Mineral Industry*, Annual issues.

G. P. Grimsley, Michigan Geol. Survey, vol. 9, pt. 2, 1904.

G. P. Grimsley and E. H. S. Bailey, Kansas Geol. Survey, vol. 5, 1899.

E. C. Eckel, Cements, limes, and plasters, 2d ed., New York, 1907.

G. I. Adams, Gypsum deposits of the United States, *Bull.* 223, U. S. Geol. Survey, 1904.

H. Ries, Economic geology, New York, 1916, pp. 244–259 (with references).

R. C. Wallace, Gypsum and anhydrite in genetic relationship, *Geol. Mag.*, vol. 1, 1914, pp. 271–276.

D. H. Newland and H. Leighton, *Bull.* 143, N. Y. State Mus., 1910.

A. F. Rogers, Notes on the occurrence of anhydrite in the United States, School of Mines Quarterly, vol. 36, 1915, pp. 123–142.

slowly alters to gypsum, and many occurrences of apparently solid gypsum contain remnants of anhydrite.

Beds of gypsum and anhydrite occur in many water laid formations all over the world. Usually gypsum predominates but alternating beds of the two are common. Beds of anhydrite up to 300 feet in thickness are found in the Permian of central Germany in connection with the potash salts (p. 312).

Anhydrite is not abundant in the United States but occurs with gypsum overlying thick salt beds in Louisiana and southern Texas; beds of anhydrite are also known from southern California, Nevada, Kansas, Nova Scotia, and New Brunswick.

The gypsum beds of the United States are rarely more than 30 or 40 feet thick though there may be several in any one section. They are interstratified with limestone or shale; in places they are of great purity and snow white; but frequently gypsum is also intergrown or interbedded with thin streaks of shale or limestone. The compact, translucent variety is called alabaster and is used for ornamental objects; gypsum in larger plates or crystals is called selenite. Recent surface deposits, mixed with clay are known as "gypsite."

A remarkable series of gypsum beds, in part alternating with anhydrite have been described from the pre-Cambrian(?) of the Palen Mountains[1] in southern California. Economically important gypsum deposits are found in the Salina (Silurian) formation in northern New York and extend parallel to the south shore of Lake Ontario.

Gypsum beds are also extensively worked in Michigan where they are of Mississippian age (Lower Carboniferous). Equally important beds of the same age are exploited in Nova Scotia and New Brunswick. Iowa, Kansas, Texas, Oklahoma, New Mexico and other states are rich in gypsum of Permian age; in the western part of this region gypsum occurs at several horizons in the "Red Beds" whose age ranges from upper Carboniferous to Jurassic.

Exceptionally thick, but not easily utilized deposits of uncertain age overlie the "salt domes" (p. 310) of Louisiana and Texas. Tertiary deposits are known from California and Quaternary "gypsite" is abundant in Kansas, Oklahoma and Texas. The Tertiary beds in the basin of Paris, France, are rich in gypsum, hence the name "Plaster of Paris."

[1] E. C. Harder, *Bull.* 430, U. S. Geol. Survey, 1910, pp. 407–416.

Uses.—Gypsum finds extensive use in various industries. Ground in its natural state, it is employed as a fertilizer (land plaster), to counteract alkali in soils, to retard the setting of cement and for numerous chemical purposes. It is often used as a "filler" or adulterant. Most important is, however, its use as structural material. For this purpose it is calcined at 350° F. when a large part of the water is expelled. After grinding and mixing with water gypsum forms again and the whole sets to a hard mass called stucco or plaster of Paris. The use of gypsum is increasing rapidly. In 1917 the production in the United States was 2,700,000 tons.

Stability and Solubility.—As noted above gypsum is transformed to anhydrite in sea water at 27° C. In pure water it begins to change slowly to anhydrite at 66° C. At or above 27° C., a temperature often reached in salt lakes in tropical countries alternating beds of gypsum and anhydrite may form, as indeed is often observed. In nature both hydration and dehydration takes place but the changes are very slow.

The solubility of gypsum is a complicated problem owing to the existence of metastable forms—the hemi-hydrate and the soluble anhydrite—and it has only lately been worked out by van't Hoff and Meyerhoffer.[1] The solubility of gypsum in water reaches a maximum of 0.21 per cent. at 40° C., and decreases slightly above this temperature. At 66° C. the solubility of anhydrite is, of course, equal to that of anhydrite, but beyond this point it decreases rapidly so that at 100° C. it is 0.06 per cent. and at 200° C. only about 0.005 per cent.[2] Other calcium salts, having a common ion, depress the solubility of gypsum, but sodium chloride increases it about three times owing to formation of $CaCl_2$, so that a saturated solution of NaCl can hold 0.54 per cent. at $CaSO_4$ at 23° C. and 0.75 per cent. at 82° C.

SODIUM SULPHATE AND SODIUM CARBONATE

Occurrence.—Most of the soda of commerce is an artificial product from common salt, but both the carbonate and the sulphate of sodium are often contained in saline desert lakes or in residues from such lakes. The ordinary white efflorescence on the *playas* of the deserts consists of these salts together with more or less sodium chloride and a little of the chlorides and

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[1] Summarized by Cameron and Bell, *Bull.* 33, U. S. Bureau of Soils, 1906.

[2] A. C. Melcher, *Jour.* Am. Chem. Soc., vol. 32, 1910, pp. 50–66.

sulphates of potassium and magnesium; the soda lakes contain all these salts.

In the United States the commercial utilization has been attempted at Owens Lake, in California, at the Ragtown lakes, in Nevada, and at the Wyoming soda lakes.

T. M. Chatard's[1] work on Owens Lake, where sodium carbonate forms a little over one-third and sodium sulphate about one-seventh of the dissolved salts, showed that the order of deposition upon evaporation is: (1) trona ($Na_2CO_3.NaHCO_3.2H_2O$); (2) sodium sulphate; (3) sodium chloride, and (4) the easily soluble normal sodium carbonate. The deposits at Ragtown are even richer in carbonate of soda,[2] but the evaporation by solar heat did not prove successful as a commercial process. One or two of the Wyoming lake deposits are rich in soda. At Green River borings in the Wasatch sandstone (Eocene?) at depths of 125 and 700 feet disclosed well water forming an almost concentrated solution of sodium carbonate, which for a time was utilized for the manufacture of caustic soda; the process was based on reaction with caustic lime.

The alkali lakes in the arid regions of Wyoming[3] and New Mexico form deposits leached from surrounding Mesozoic and Cenozoic sediments. The thickness of the salt beds amounts to 15 feet at most and they extend over as much as 100 acres. The salts consist mainly of mirabilite, epsomite, natron, and halite.

Sodium sulphate is much more soluble in warm than in cold water, but as the similar variation for sodium chloride is small, "a mere change of temperature between summer and winter in salt lakes may cause mirabilite ($Na_2SO_4.10H_2O$) to separate out or to redissolve." The Great Salt Lake, according to Gilbert, deposits sodium sulphate during winter.

SODIUM NITRATE

The alkaline nitrates are very soluble salts which are found in larger masses only under exceptional conditions. Sodium nitrate is present in the soil and is produced by the so-called

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[1] T. M. Chatard, Natural soda, *Bull.* 60, U. S. Geol. Survey, 1890.

[2] F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 238.

[3] A. R. Schultz, Deposits of sodium salts in Wyoming, *Bull.* 430, U. S. Geol. Survey, 1910, pp. 570–589.

W. C. Knight and E. E. Slosson, Alkali lakes and deposits, *Bull.*, Wyo. Exper. Sta., 1901, p. 49.

nitrifying bacteria[1] or by reactions between organic nitrogenous matter and alkaline salts. Sodium and potassium nitrates of great purity are sometimes found as efflorescences and veinlets on sheltered cliffs of various rocks and in caves and are in many cases produced by organic agencies. Calcium nitrate is known from limestone caves. Naturally, nitrate deposits are most common in arid countries. In minor quantities nitrates are widely scattered in the Western States and very frequently they are associated with volcanic rocks,[2] particularly rhyolite but also tuffs, basalts and lake beds in regions of volcanic activity. The volcanic origin of these nitrates is not accepted by all writers but nevertheless it is the most probable theory advanced.

There are two sources of nitrogen which may be utilized by nature for the development of ammonia salts and nitrates. 1. The nitrogen in the air, which may be fixed by organisms or by electric atmospheric discharges and entrainment in rain water. 2. The nitrogen from the interior of the earth, which possibly is contained in the magma as a nitride of boron or of some metal. At any rate the volcanic gases and exhalations frequently contain nitrogen and ammonia; it is held by many that a fixation of nitrogen from this source as nitrates is well possible.[3]

The only place where nitrates are present in abundance is in the Atacama desert in northern Chile.[4] These wonderful deposits

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[1] H. S. Gale, Nitrate deposits, *Bull.* 523, U. S. Geol. Survey, 1912.

[2] G. R. Mansfield, Nitrate deposits in southern Idaho and eastern Oregon, *Bull.* 620, U. S. Geol. Survey, 1916, pp. 19–44.

Whitman Cross, *Am. Jour. Sci.*, 4th ser., vol. 4, 1897, p. 118.

W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905, p. 121.

C. DeKalb, *Min. and Sci. Press*, May 6, 1916.

[3] F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 256.

[4] The literature is very extensive and only part can be quoted.

L. Darapsky, Das Departement Tal-tal, Berlin, 1900. Ref. Zeitschr. prakt. Geol., 1902, p. 153.

R. A. F. Penrose, *Jour. Geol.*, vol. 18, 1910, pp. 1–32.

S. H. Loram, *Min. and Sci. Press*, Jan. 15 and 29, 1910.

W. F. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, pp. 253–259.

L. W. Strauss, *Min. and Sci. Press*, June 13 and 20, 1914.

J. T. Singewald, Jr., and B. L. Miller, *Econ. Geol.*, vol. 11, 1916, pp. 103–114.

Lorenzo Sundt, *Econ. Geol.*, vol. 12, 1917, p. 89.

A. H. Rogers and H. R. Van Wagenen., The Chilean Nitrate Industry, *Bull.* 134, Am. Inst. Min. Eng., Feb., 1918, pp. 505–522. Discussion, *Bull.* 136, pp. 845–848.

S. H. Salisbury, Jr., *Mineral Industry*, Annual issues.

practically supply the world with nitrates; the annual production now (1916) amounts to nearly 3,000,000 metric tons. The deposits are situated in the provinces of Tarapaca and Antofagasta in the interior dry valleys between the Coast Range and the Andes, at elevations ranging from 1,000 feet to 3,000 feet, and they extend for 300 miles parallel to the coast. The lowest depressions are often occupied by salt flats with a little nitrate and, in the higher region, by borax flats. The nitrate deposits lie on the gentle slopes of the valleys. The nitrate bed is a superficial formation of considerable though irregular extent; it lies below an overburden of a few feet of loose crumbly material with subangular gravel, becoming harder toward the bottom. This overburden contains some nitrate and often much sodium chloride, sodium sulphate and gypsum, as well as a little sodium iodate. The "Caliche" or nitrate bed is a reddish brown sandy gravel cemented with salts; it averages a few feet in thickness. Below the "caliche" lies rudely stratified sand, gravel or clay, often of considerable thickness. The "caliche" averages about 25 per cent. sodium nitrate and the lower limit of workable material is placed at 15 per cent. Associated with the nitrate are a large amount of sodium chloride, more or less of the sulphate and borates or calcium and sodium, and a small but constant quantity of sodium iodate. Small quantities of the nitrates of potassium, calcium and barium as well as a little calcium iodate and iodo-chromate (lautarite and dietzeite) are found. Very curious is the occurrence of a small amount of sodium perchlorate.

The material mined is usually of the following composition:

	Per cent.
Sodium nitrate	14–25
Potassium nitrate	2– 3
Sodium chloride	8–50
Sodium sulphate	2–12
Calcium sulphate	2– 6
Magnesium sulphate	0– 3
Sodium biborate	1– 3
Sodium iodate	0.05– 1
Sodium perchlorate	0.1 – 0.5
Insoluble	0–50

The origin of the nitrate deposits of Chile is a much debated question and few authors are in agreement

The theory advanced many years ago by Pissis, the Chilean geologist, and followed lately, for instance, by Rogers and Van Wagenen accounts for the deposits by fixation of atmospheric nitrogen by thunderstorms and its descent from the Andes in the underground circulation and ascent to the surface by capillarity.

Penrose and others hold that the nitrate came from beds of bird guano accumulated at the time when the Coast Range did not exist and that the nitrates were gradually leached and mingled with the salt waters of a closed basin. Others are inclined to consider the deposits caused by ordinary bacterial fixation or by oxidation of nitrogenous vegetable matter. Singewald and Miller think that the nitrates have been carried down by the ground water and emphasize that only the usual processes in operation everywhere, have been active. The accumulation is simply caused by the abnormally dry climate.

All these explanations appear inadequate or forced. The nitrate deposits are not marine or lacustrine. Their extent corresponds in a most remarkable way to the Jurassic and Cretaceous tuffs and lava flows which occupy so much space in this region and the conclusion is inevitable that there must be some causal connection.[1] It is probable that the nitrates are of volcanic origin and that the nitrogen was contained in the rocks mentioned from which they have been leached under unusual climatic conditions. This view of the origin is also supported by F. W. Clarke.[2] The constant presence of borates is an additional suggestive fact.

The world's need of iodine is now supplied by the nitrate region of Chile. The production was 709,000 kilograms in 1915.

BORATES[3]

General Occurrence.—Borates and other boron compounds appear in nature under conditions indicating widely differing modes of origin. As complex and insoluble borosilicates like

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[1] Unpublished observations by W. L. Whitehead and W. Lindgren.

[2] Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 258.

[3] G. E. Bailey, The saline deposits of California, *Bull.* 24, California State Min. Bur., 1902.

M. R. Campbell, Reconnaissance of the borax deposits of Death Valley and the Mohave Desert, *Bull.* 200, U. S. Geol. Survey, 1902.

C. R. Keyes, Borax deposits of the United States, *Trans.*, Am. Inst. Min. Eng., vol. 40, 1909, pp. 674–710.

tourmaline and datolite they are disseminated in igneous and metamorphic rocks or in pegmatite dikes and fissure veins, but are here of no economic importance except that tourmaline occurring in this manner is sometimes utilized as a gem stone. As boric acid and borates of calcium and magnesium they appear in volcanic exhalations, of which the most famous are the "soffioni" of Tuscany, Italy, from which large amounts of boric acid have been recovered. Borates, principally in the form of borax ($Na_2B_4O_7.10H_2O$) occur in hot springs and in lakes of volcanic regions. Borax was first obtained from such lakes situated in Tibet. Von Schlagintweit reports it as a hot-spring deposit in the province of Ladak, on the headwaters of the Indus. According to A. Forbes a calcium borate is being deposited at the hot springs of Baños del Toro, Chile. The thermal waters of the California Coast Ranges and Nevada (page 61) often contain boron, sometimes in large quantities. The borates from these springs are sometimes accumulated in little lake basins and there deposited by evaporation as borax crystals. About 40 years ago much borax was won from the Borax Lake, Lake County, California. The evaporated salts contained 62 per cent. sodium carbonate, 20 per cent. sodium chloride, and 18 per cent. borax.

The borates occur abundantly in the *playas*, or shallow basins intermittently covered by water, in southern California, Nevada, Oregon, Argentina, and Chile; the salts are borax, ulexite ($CaNaB_5O_9.8H_2O$), and colemanite ($Ca_2N_6O_{11}.5H_2O$).

The Quaternary borax beds are probably derived from leaching of deposits of colemanite in Tertiary lake beds, formed during volcanic epochs. The deposits in southern California now furnish most of the world's boron salts.

Finally, boron is contained in sea water and appears in small quantities in the form of magnesium borates, principally boracite ($Mg_7Cl_2N_{16}O_{30}$), in saline residues.

Marine Borate Deposits.—The marine deposits are mainly confined to the beds of potassium salts in central Germany, but

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J. H. van't Hoff, Zur Bildung der ozeanischen Salzablagerungen, 1905.

General: F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, pp. 243–253.

C. G. Yale and H. S. Gale, in *Mineral Resources*, U. S. Geol. Survey, annual publication.

H. S. Gale, The origin of colemanite deposits, *Prof. Paper* 85, U. S. Geol. Survey, 1913, pp. 3–9.

FIG. 97.—Sketch map showing distributions of borate deposits in California and Nevada. *After H. S. Gale, U. S. Geol. Survey.*

borates have also been observed in sodium chloride, anhydrite, or gypsum. The principal occurrence is as boracite and several other rare borates in the carnallite region (page 313)—that is, in the deposits of the last mother liquors of evaporating sea water. The boracite usually forms small crystals or concretions, but one occurrence is recorded of a mass weighing about 1,400 pounds. A few hundred tons of borates are annually obtained by crystallization. The boron compounds then remained with the most easily soluble salts and were finally precipitated as a magnesium salt because of the predominance of that metal over calcium in the sea water.

Borax Marshes.—The deserts of San Bernardino and Inyo counties in California, and also those of Nevada and Oregon, are rich in borate deposits (Fig. 97). The desolate plains between the barren ranges contain many shallow basins, which at times, during the brief seasons of rain, are covered with thin sheets of water. The evaporation of this water leaves dazzling expanses of white salt deposit or efflorescence, some of which may become covered by the fine sand carried by the desert storms. These deposits were discovered about 1870 and for many years yielded a large production of borax at Searles Marsh (60 miles north of Barstow), Death Valley, and other places. Though enormous quantities of these salts remain they are now of little or no importance; the richer and more easily worked colemanite deposits have replaced them. The crusts are rarely more than 1 foot thick, the percentage of borax varying considerably. According to Bailey, the crude salt from Searles Marsh yielded 50 per cent. sand, 12 per cent. sodium chloride, 10 per cent. sodium carbonate, 16 per cent. sodium sulphate, and 12 per cent. borax. Borings showed 20 feet of clay and sand with crystals of calcium borate, underlain by a bed of solid trona 28 feet thick, and below this 350 feet of clays impregnated with hydrogen sulphide. On ground that had been worked over a new crust formed by capillary action that was thick enough to remove in 3 or 4 years. The area productive of borax amounts to about 1,700 acres, slightly depressed below the general level of the *playa*, on which in wet seasons stands about 1 foot of water.

Though no ulexite was found at Searles Marsh, it is common in many other *playa* deposits, both in California and Nevada and in Argentina; it usually forms concretions of silky fiber, known as "cotton balls."

Tertiary Lake Beds.—The borates in the marshes and *playas* have undoubtedly been leached from the older deposits in the Tertiary lake beds, which have been recognized at many points in Inyo, San Bernardino, Kern, Los Angeles, and Ventura counties, California. These beds yield colemanite almost exclusively and it is evident that the borax and ulexite of the marshes are mainly products of secondary reactions of the leached colemanite with the sodium salts of the *playas*.

These colemanite deposits begin near the Pacific coast at Piru, Ventura County, and near Saugus, Los Angeles County, where, according to Keyes, they lie in a series of yellow clays and sandstones probably Miocene in age and several thousand feet in thickness. The mineral is present as nodules in clay, and above the borate beds are strata of gypsum.

Fig. 98.—Lila C. borate mine, Inyo County, California, closed in 1915.

Other important beds are in the foot-hills of the Calico Mountains north of the Mojave River; they have a steep dip and have been mined by shafts to a depth of 400 feet. These large masses of low-grade colemanite shales, with 7 to 20 per cent. boric acid, are not mined now, attention being confined to two solid beds of the mineral 7 to 10 feet thick. Rhyolite tuffs lie underneath the borate beds.

The richest colemanite beds are, however, in the Furnace Creek region of the Funeral Range, which overlooks Death Valley, in Inyo County (Fig. 98). About 4,000 feet of Tertiary non-fossiliferous sediments are recognized here, which form a broad belt obliquely crossing the range and dipping 20° to 45° N. E. The lower and thicker part, according to Keyes, consists of conglomerates and sandstones, above which are olive-colored clays interbedded with basalts. The upper part of the clay series carries gypsum, colemanite, and thin layers of limestone. The borate beds are traceable for 25 miles. Within the colemanite-bearing beds, which may be as much as 50 feet thick, the bluish clays are thickly interspersed with milky white layers or nodules of the white coarsely crystalline mineral, mingled with more or less gypsum, according to Keyes. The solid layers may be 15 feet thick. Near by the clays are impregnated with fine particles of colemanite and yield 10 to 25 per cent. boric acid, but these low-grade deposits are not utilized at present.

Production and Uses.—The borate industry is now concentrated in southeastern California and has shown great expansion in the last years. In 1916, 103,000 tons of crude colemanite were mined, averaging about 25 per cent. B_2O_3. The richest mineral is hand-sorted, and the poorer grades are milled, roasted, and screened, the last process effecting the separation of the colemanite from the gangue. The products are shipped direct to the sea board, where the material is manufactured into borax and boric acid. The further treatment involves boiling with soda for the manufacture of borax or with sulphuric acid if boric acid is desired. Under the influence of these new discoveries the price of borax has gradually decreased; in 1916 it was from 6 to 8 cents per pound.

Borax is extensively used in industrial chemistry, in metal enameling, in medicine and in the household.

Origin.—The colemanite deposits which in places occur with gypsum and limestone are surely not of marine origin and can hardly be supposed to be saline precipitates from evaporating lake water. The mode of occurrence in specimens suggests replacement and it has indeed been shown lately by H. S. Gale[1] that some colemanite deposits are of epigenetic nature. The replacing boron solutions may have been leached from the sediments or they may have ascending waters in genetic connection with basalt flows formed in close connection with the beds. The strontianite deposits found in similar lake beds (p. 380) form a somewhat analogous occurrence. Further investigations of the colemanite deposits is highly desirable.

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[1] *Prof. Paper* 85, U. S. Geol. Survey, 1913, pp. 3–9.

Colemanite and the allied species priceite have been formed in recent borax marshes, for instance, at Searles lake and would undoubtedly be formed from alkaline borate solutions in contact with calcite or calcium carbonate waters.

Van't Hoff[1] has produced pandermite ($Ca_8B_{20}O_{38}.15H_2O$) and colemanite from the heptaborate ($Ca_2B_6O_{11}.7H_2O$) and states that ulexite, pandermite, and probably colemanite can be formed at temperatures from 25° C. upward. Regarding the separation of borates he states that while the first salts precipitated in oceanic waters are calcium salts, different relations exist with the borates; for these the saturation point is not reached until carnallite is precipitated.

SODIUM CHLORIDE

Occurrence.[2]—Sodium chloride or common salt forms beds in sedimentary rocks and in most cases its derivation by evaporation of saline solutions is clear. Only a small part of the four million tons of salt produced in the United States is mined in solid form. Most of it is obtained from brines derived from solution of salt beds by natural waters or by water forced down into bore-holes to the saline strata; much also is produced by evaporation of sea water or water of saline lakes, such as the Great Salt Lake of Utah.

Salt beds are present in formations of different ages, but are perhaps most common in the Permian and Triassic strata; the oldest saline rocks in the United States are those of the Silurian in New York State. As may be easily understood from the general statements on previous pages, strata of calcium sulphate are ordinarily associated with salt beds and should appear below them; owing to recurrent and shifting epochs of desiccation the order may be reversed and gypsum beds appear above the salt. It is also very common to find crystals or streaks of anhydrite or gypsum with salt, as well as streaks of clay. In thickness

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[1] J. H. van't Hoff, Zur Bildung der ozeanischen Salzablagerungen, 2, 1909, pp. 45–75.

[2] U. S. Geol. Survey, *Mineral Resources* (annual reports). See especially W. C. Phalen, *Mineral Resources*, 1907-1911.

Mineral Industry (annual issues).

G. D. Harris, *Bull.* 7, Louisiana Geol. Survey, 1908.

J. O. von Buschman, Das Salz, Leipzig, 1906 and 1909, 2 volumes.

W. C. Phalen, Technology of salt making in the United States, *Bull.* 146, U. S. Bureau of Mines, 1917 (with description of deposits).

salt beds vary enormously—from the thinnest strata to masses 1,500 feet or even more in depth. A bore-hole near Speerenberg, in the German potash region, penetrated 3,900 feet of

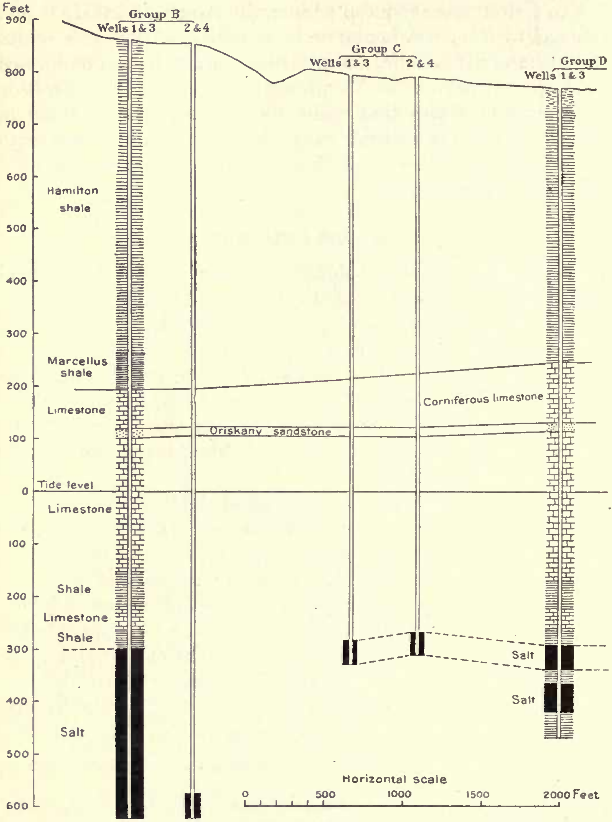

FIG. 99.—Sections of salt wells, Tully, New York. *After F. J. H. Merrill.*

salt, but here, as in so many other places, the apparent thickness may be deceptive, being due to movements of folding and faulting. Besides, the plasticity of salt is remarkably great, much greater than that of the accompanying clays and anhydrite, and this, as the German geologists have found, leads to most astonishing and confusing stratigraphic relations.

From the calcium sulphate secondary sulphur often results and may form thick beds. Hydrocarbons and carbon dioxide are often contained as inclusions in the salt. The difficulties of accounting for the great thickness of salt beds have already been referred to. It is evident that by evaporation of sea water with 3.5 per cent. salt in a basin 100 meters deep a bed less than 2 meters of salt would accumulate. The theories of Ochsenius and Walther attempt to explain this in different ways, as described on previous pages. The special conditions in Louisiana will be referred to later.

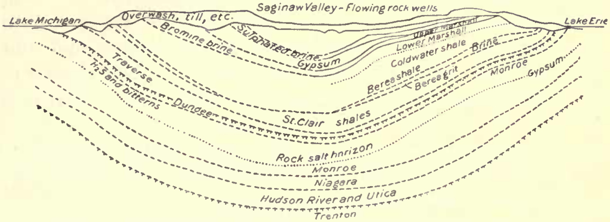

FIG. 100.—Section of lower Michigan basin. *After A. C. Lane.*

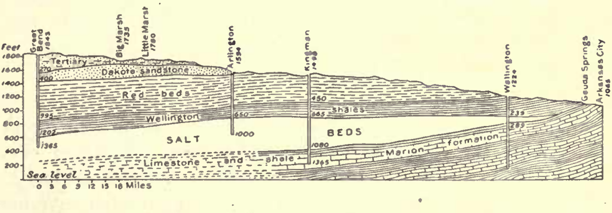

FIG. 101.—Section of Permian salt formation in Kansas. *From Mineral resources of Kansas.*

Examples.—Salt beds occur in New York State in the red Salina shales of the Silurian and underlie a considerable area. Much of the salt is recovered from artificial brines. The salt forms pure lenticular masses and layers interbedded with soft shales, limestone, and gypsum, the salt-bearing formation having a variable thickness up to 470 feet (Fig. 99). At Ithaca several beds of salt occur at a depth of 2,244 feet with a total thickness of 248 feet. A magnesian limestone, containing gypsum, lies above. Rock salt is mined at several places, one shaft lately opened at Cuylerville being 1,100 feet deep and reaching a salt bed 21 feet in thickness. Salt has also been mined near the outcrops of the beds at Livonia. Similar beds are worked in Ohio by bore-holes and brines.

The greatest salt production in the United States is derived from Michigan.[1] The salt occurs as large beds at different horizons in the Salina formation and also in the sandstones of the Mississippian or Lower Carboniferous (Fig. 100). The salt is recovered by means of natural and artificial brines; bromine, in which these brines are unusually rich, is recovered as a by-product of the final mother liquor. Deep mining has been undertaken under considerable difficulties near Detroit.

Kansas is likewise among the great producers.[2] Some salt is obtained from salt springs in the Carboniferous and on the "salt plains" leached from Permian beds. From the latter the principal product is derived; it occurs interstratified with shales, the total thickness of the salt beds being at most 500 feet. Some of the beds are said to be over 200 feet thick, but generally they are much less (Fig. 101).

In the western arid States *playa* deposits of salt are common in the dry basins between the ranges; they are usually thin, though at Danby, in southern California, there are solid beds 22 feet in thickness, according to Bailey.[3]

The most noted deposit is that of Salton, Imperial County, where the basin lies below the level of the sea. Before the recent flooding by the Colorado River an important production was maintained here. A large area is covered by salt crusts 10 to 20 inches in thickness. Below this lies a thin mud deposit covering another salt crust. Deeper borings encountered 22 feet of black mud containing salt and soda, and this covers 270 feet of hard clay.[4]

The desert regions of northern Africa and central Asia offer similar occurrences in abundance.

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[1] A. C. Lane, *Mineral Industry*, vol. 16, 1907; also vol. 19, 1910.
A. C. Lane, *Water-Supply Paper* 30, U. S. Geol. Survey, 1899.

[2] M. Z. Kirk, Mineral resources of Kansas, Univ. Geol. Survey, 1898.

[3] G. E. Bailey, *Bull.* 24, California State Min. Bur., 1902, p. 128.

[4] G. E. Bailey, *idem*, p. 126.

Large deposits of impure salt mixed with clay have been worked for a long time in the Alpine Triassic of Tyrol; they lie between limestone beds. Another important saline region fringes the outside of the Carpathian chain in Roumania, Transylvania, and Galicia and is contained in Miocene sands and clays. The beds are generally greatly disturbed, brecciated, and pressed. The best-known place where mining is carried on is the celebrated mine of Wieliczka, in Galicia, now about 1,000 feet deep, which is much visited by tourists on account of the picturesque and extensive workings with elaborate carvings in solid salt. The salt beds of the Stassfurt region will be described later.

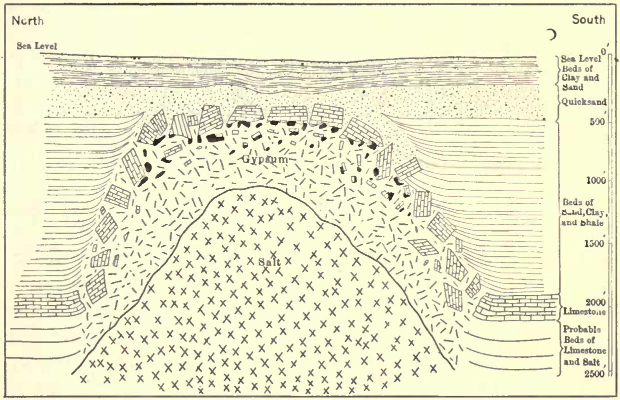

FIG. 102.—Vertical section of salt dome, based on borings at Calcasieu parish, Louisiana. Black areas represent sulphur. *After Kirby Thomas.*

The Salt Deposits of the Gulf Coast.[1]—The greatest salt deposits in the United States have been discovered by borings in

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[1] G. D. Harris, *Bull.* 7, Geol. Survey Louisiana, 1908.

A. F. Lucas, The possible existence of deep-seated oil deposits on the Gulf Coast, *Bull.* 139, Am. Inst. Min. Eng., 1918, pp. 1119–1134.

G. S. Rogers, Intrusive origin of the Gulf Coast Salt Domes, *Econ. Geol.*, vol. 13, 1918, pp. 447–485.

E. R. DeGolyer, The theory of volcanic origin of the salt domes, *Bull.* 137, Am. Inst. Min. Eng., 1918, pp. 987–1000.

Louisiana and the adjoining coast belt of Texas; they show many unusual features and some difficulty has been experienced in explaining their genesis. Above the low and swampy coast west of New Orleans rise a number of low mounds or knolls and below these most of the salt has been found. It does not occur in regular beds, but as enormous subterranean domes, surrounded on all sides by thick and often steeply dipping beds of Quaternary and Tertiary clays and sands. At some places a thin-bedded Cretaceous limestone appears at the surface. Fig. 102 gives a suggestion of the strange relations encountered. At Petite Anse, according to Harris, the drill shows 2,263 feet of almost pure salt, followed by 70 feet of foreign matter, below which the drill again enters rock salt of unknown thickness. On Côte Carline the drill entered salt at 334 feet and continued in salt without change till the drilling ceased at 2,090 feet. At Belle Isle[1] the Knapp Well No. 1 penetrated 2,000 feet of salt and, below this, 236 feet of anhydrite and gypsum. Another well at Humble is said to have penetrated 5,410 feet of salt.

Oil, gas, and sulphur are often met in the drill-holes. Gypsum and anhydrite, in beds 200, 400, or even 600 feet thick cover the salt in some places, or again the salt may be overlain (as at Spindletop, Texas) by several hundred feet of a porous limestone carrying oil. The dip of the loose strata forming the outside of the dome is steep and bore-holes only a short distance from those disclosing salt may fail to encounter it. Naturally the published sections, based on a few bore-holes, are more or less problematical as to structure.

These enormous salt resources are as yet little utilized. Rock salt was mined in 1915 at Weeks Island, where the shaft is 645 feet deep, and at Avery Island, at a depth of about 500 feet. The shafts are sunk in heavy, wet ground until the impermeable salt is reached. In places there is considerable danger of flooding the mine by driving into the loose strata.

According to R. T. Hill, these wonderful salt domes are deposited by ascending solutions; the uplift of surrounding strata is caused by the hydrostatic pressure of salt solutions and oil rising through fissured rocks. According to L. Hager and A. C. Veatch, the domes are uplifts caused by laccolithic intrusions. According to G. D. Harris, the uplifts are produced by the expanding power of growing salt crystals, the concentrated

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[1] A. F. Lucas, *Trans.* Am. Inst. Min. Eng., vol. 57, 1917, p. 1034.

solutions rising at the intersections of fissures; the salt is derived from underlying Paleozoic or Mesozoic beds.

None of these views are fully convincing. A brief consideration of the relations of solubility of sodium chloride will show that only a very small quantity could have been precipitated as the temperature of the ascending solutions was lowered and that, therefore, the quantity of primary salt required by Harris' hypothesis would be incredibly large.

Recent literature has shown the existence of many salt domes along the coast and some quite a distance inland.[1] The same kind of salt domes are also found on the isthmus of Tehuantepec back of Puerto Mexico,[2] accompanied in places by oil and gas. The foreign literature indicates that such salt domes also exist in northern Germany and in Transylvania.[3]

Hopkins shows clearly that the salt dome at Palestine, Texas, is caused by a highly localized vertical uplift of quaquaversal form. Rogers and DeGolyer arrive at similar conclusions. It is probable that these domes are Permian or Carboniferous salt beds forced up through the softer sediments. This is made possible by the extraordinary plasticity of rock salt, which easily yields to deformation. The nature of the force producing these uplifts remains in doubt.

Composition, Production and Use.—Rock salt is usually very pure aside from the occasionally occurring admixture of clay, the tenor in NaCl ranging from 96 to 99 per cent. Calcium sulphate is the principal impurity and is often present to the amount of over 1 per cent. Salt from some desert lakes contains sodium carbonate and sulphate. In 1917, 1,605,000 short tons of rock salt was mined in the United States. The total production of salt for the same year was nearly 7,000,000 short tons. The average price was $2.86 per ton.

The wide range of uses of salt for culinary, preservative, and industrial purposes need not be specified. Large amounts are used in the manufacture of other sodium salts, particularly the carbonate.

In 1917, 450 tons of bromine were produced in Michigan, Ohio, and West Virginia, the normal price is 50 cents per pound.

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[1] O. B. Hopkins, *Bull.* 616, U. S. Geol. Survey, 1917, p. 28.

[2] Burton Hartley, *Econ. Geol.*, vol. 12, 1917, pp. 581–588.

[3] F. F. Hahn, *Econ. Geol.*, vol. 7, 1912, pp. 120–135.

THE GERMAN POTASSIUM SALTS[1]

If carried to its conclusion the process of evaporation of sea water will result in the deposition of the easily soluble chlorides and sulphates of potassium and magnesium, also chloride of calcium. Evidently this seldom takes place, in part for the reason given on page 292. Almost the only locality thus far discovered where the whole sequence of salts is present is in central Germany, north and south of the Harz Mountains, in formations of Permian age. These deposits are now mined by about fifty companies and yielded in 1913 about twelve million tons of potassium salts, of which 85 per cent. is used as soil fertilizer and the remainder for general industrial purposes. The value of this production is about $45,000,000. The mining is done exclusively by shafts from 1,000 to 2,500 feet deep. Circular shafts lined with concrete or iron tubing are used and the greatest caution is necessary to prevent influx of water during sinking or working; if the water once breaks in, the mine will probably have to be abandoned. The old workings are filled with waste or rock salt.

Some of the products are sold for fertilizers without further chemical treatment. Carnallite is the most important of these, and next to it comes kainite; as mined, both are mixed with 30 or 40 per cent. of common salt. Other chemical products from carnallite and other salts are chloride and sulphate of potassium and potassium-magnesium sulphate. Kieserite is refined to magnesium sulphate.

The larger part of the bromine production of the world is also obtained from the mother liquor resulting from the solution of the Stassfurt salts. The remaining part is derived from the brines of Michigan.

The potassium salts lie as a relatively thin series of strata over a heavy bed of rock salt in the Permian and are in turn covered by Triassic sandstones and limestones, and finally by the Tertiary and Quaternary beds. They form a series of

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[1] H. Precht, Die Salz industrie von Stassfurt, 1889.

R. Ehrhardt, Die Kali-Industrie, 1907.

E. Pfeiffer, Handbuch der Kali-Industrie, 1887.

Beyschlag, Everding, Erdmann, Loewe, and Paxmann, Deutschlands Kalibergbau, 1907.

Summaries: F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, pp. 221–228; G. P. Merrill, Non-metallic minerals, 1905; R. Meeks, *Mineral Industry*, 1906.

faulted synclines and anticlines, in places approaching closely to the surface, but here generally changed by secondary leaching processes (Fig. 103).

The general section is as follows,[1] counted from the bottom of the Triassic sandstone.

No.	Thickness in meters	Character of strata
1	20 to 30........	Red clay with a little anhydrite and rock salt.
2	50..............	Rock salt.
3	1 to 5..........	Anhydrite with salt.
4	5 to 15.........	Red clay with anhydrite and rock salt.
5	100 to 150......	Younger rock salt.
6	40 to 90........	Main anhydrite.
7	4 to 10.........	Salt clay.
8	30 to 40........	Carnallite zone. Carnallite ($KCl.MgCl_2+6H_2O$).
9	20 to 40........	Kieserite zone. Kieserite ($MgSO_4+H_2O$).
10	40 to 60........	Polyhalite zone. Polyhalite ($2CaSO_4.MgSO_4.K_2SO_4+2H_2O$).
11	300 to 500......	Older rock salt, with narrow streaks of anhydrite, interpreted as annual deposits.
12	70 to 100.......	Older anhydrite.
13	4 to 10.........	Limestone (Zechstein Kalk). Marine deposit.
14	0.5 to 4........	Black copper-bearing shale. Conglomerate.
		Lower Permian and Carboniferous beds.

The older series (Nos. 8 to 11) closed with deposition of potassium salts. The younger series (Nos. 1 to 7) contains no potassium salts.

Rock salt is really present throughout the section, for the carnallite zone, which yields the greatest quantity of crude product, contains only about 55 per cent. of carnallite, including also 25 per cent. rock salt and 16 per cent. kieserite. The kieserite zone yields 65 per cent. rock salt and only 17 per cent. kieserite. Rock salt, kieserite, sylvite (KCl), carnallite ($KCl.MgCl_2+6H_2O$), and kainite ($KCl.MgSO_4+3H_2O$) are the main products.

The German geologists have shown that extensive secondary changes have taken place in the salt beds—in part immediately after the deposition, in part much later, following the Triassic sedimentation and progressing even now. These post-Triassic

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[1] H. Everding, Deutschlands Kalibergbau, 1907, p. 36.

changes have occurred both in the croppings and at greater depth. The minerals just enumerated may occur in all three generations, but in addition a large number of more or less complicated secondary compounds were formed.

Kainite is in part a secondary product derived from carnallite by the leaching of $MgCl_2$. It forms along the crests of the anticlines. Under some circumstances a secondary mixture of potassium chloride, kieserite, and salt would be formed instead of kainite, and this constitutes an important product under the name of "hartsalz." Secondary deposits of the older type appear between the carnallite and the salt clay. In all these transformations the products are very complex.

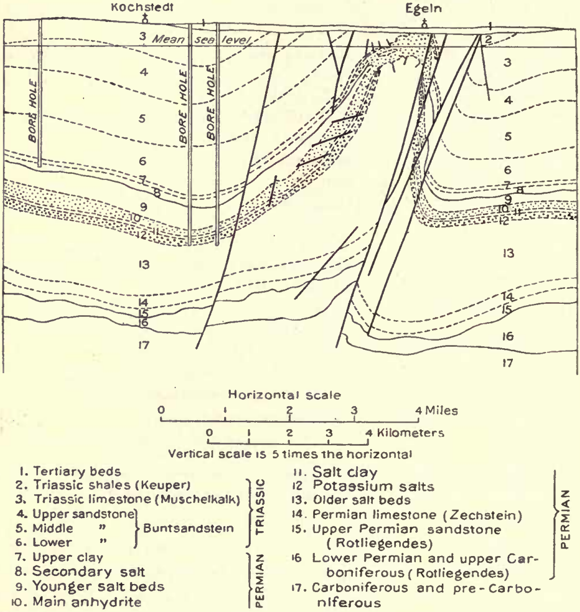

FIG. 103.—Section of the Stassfurt-Egeln anticline. *After Everding.*

Van't Hoff[1] and his associates have studied the various com-

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[1] Van't Hoff, Die ozeanischen Salzablagerungen, 1905 and 1909.

binations of salts in order to ascertain their fields of existence at temperatures ranging from 25° to 83° C. In this way they have arrived at the temperatures of stability of the various salts and consequently ascertained the minimum temperature at which they were formed. Sylvite and carnallite are stable in concentrated NaCl solutions from 0° to 85° C. Many of the rarer salts (langbeinite ($2MgSO_4.K_2SO_4$), for instance) are stable under these conditions only from 37° C. upward.

Some of the temperatures required may seem high; kieserite with sylvite, for instance, forming above 72° C. There is, indeed, some danger in using the laboratory results to explain the processes of nature, for the important element of time probably plays a considerable part in these reactions.

It has been shown by measurements, however, that the temperature in the middle depths of evaporating salt lakes is surprisingly high. Interesting results were obtained by A. V. Kaleczinsky,[1] who found the temperatures of certain Hungarian salt lakes to be as much as 71° C. during the summer at a depth of 1.3 meters, while the surface and the bottom layers were much cooler, about 20° C.

OTHER SOURCES OF POTASSIUM SALTS

The exhausted agricultural lands of all countries need potassium salts,[2] together with phosphates and nitrogen compounds. Germany is the only country in which potassium salts in easily available form occur on a large scale. The imports of the United States in 1913 amounted to about 1,800,000 metric tons while the highly stimulated domestic production since the war broke out amounts to but 20 per cent. of the former consumption. This brief statement indicates the acute situation.

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[1] A. V. Kaleczinsky, *Ref. Ann. Phys.* (4), 7, 1902.

[2] W. C. Phalen, Potash salts, *Mineral Resources*, U. S. Geol. Survey, annual publication, 1910–16 (with literature).

W. C. Phalen, Occurrence of potash salts in the bitterns of the eastern United States; *Bull.* 530, U. S. Geol. Survey, 1911.

H. S. Gale, The search for potash in the United States; *Bull.* 530, U. S. Geol. Survey, 1911. Also in *Bull.* 580, U. S. Geol. Survey, 1914, pp. 265–317.

A. R. Schultz and Whitman Cross, Potash-bearing rocks of the Leucite Hills, Wyo.; *Bull.* 512, U. S. Geol. Survey, 1912.

B. S. Butler and H. S. Gale, Alunite, *Bull.* 511, U. S. Geol. Survey, 1912.

Promising beds of potassium salts similar to those of Germany have been discovered in the Oligocene of Alsace. Other deposits occur in Tyrol, Spain, Galicia and India, but none of these have as yet contributed to the world's supply. An active search for potassium salts has been carried on in this country since 1910

Potassium in Rocks and Minerals.—Granites, pegmatites, some phonolites, and some leucite rocks contain a considerable amount of potassium, varying from 5 to 12 per cent. Unfortunately there is great difficulty in transforming the insoluble potassium silicates contained in the orthoclase, leucite, or glassy base in these rocks into soluble salts. Some pegmatite dikes, composed largely of orthoclase, yield even more than 12 per cent. of potash. If orthoclase or any potassium-bearing rock is ground to fine powder and slimed with water a certain small percentage of potash salt is converted into soluble form, probably as a potassium silicate, and it is said that such finely ground powder has some slight fertilizing power. Processes have been patented by A. S. Cushman and others based on electrolytic treatment of the slimed rock or treatment with quicklime and calcium chloride and subsequent calcining, methods by which soluble potassium salts are said to be set free. None of these processes has thus far been applied on a large scale.

Greensand marls—for instance, the Cretaceous beds in New Jersey—contain from 3 to 6 per cent. of potash besides some phosphoric acid, the former in glauconite, the latter in calcium phosphate. These marls are used in their crude state as fertilizers and the recovery in soluble form of their potassium content has been proposed, but the practical application has not as yet been attempted.

Another source of potassium has been sought in the mineral alunite, which is a hydrous potassium-aluminum sulphate of inconspicuous appearance, white or pink, compact or fine granular, rarely coarse granular. The formula of the mineral is $K_2O.3Al_2O_3.4SO_3.6H_2O$; and it contains from 8 to 11 per cent. of potash. It occurs mainly in volcanic regions, as a product of rock alteration, probably caused by waters containing sulphuric acid, and is much more common than the soluble natural alum which sometimes appears as efflorescences. The alunite is found disseminated in the rocks or in well-defined veins. Notable western occurrences are at Goldfield, Nevada; Marysvale, Utah; and the Rosita Hills, Colorado. Among the foreign deposits which have already been utilized are those at La Tolfa, in Italy; Almeria, in Spain; and Bullah Delah, in New South Wales. The transformation of alunite into soluble potassium sulphate is easily effected by calcination; part of the sulphuric acid and all of the water is volatilized, leaving soluble potassium sulphate and an insoluble residue of alumina. The manufacture of potassium sulphate from alunite began at Marysvale in 1915.

Potassium in Brines.—Potassium salts are easily soluble and therefore remain with calcium and magnesium chlorides in the last residues or mother liquors, the so-called "bitterns." Many natural brines pumped from bore-holes in salt-bearing beds contain some potassium and under favorable circumstances may be used for the recovery of these salts. Some of the Michigan brines from the Marshall formation of the lower Carboniferous (Fig. 92) contain from 3 to 5 grams per liter of potassium chloride; salt, calcium chloride, and bromine are recovered from these brines, but their potassium content appears to be too small for profitable recovery. In places certain well-defined strata yield natural brines or residual "bitterns." One such bittern from Fairport Harbor, in Ohio, on Lake Erie, contains, according to W. C. Phalen, in grams per liter, 7.4 KCl, 110.1 NaCl, 134.4 $CaCl_2$, 43.2 $MgCl_2$. Such a brine could possibly be utilized for the recovery of potassium. This stratum is almost 400 feet above the topmost salt bed from which artificial brines are pumped in Ohio.

Lakes in dry regions, especially in areas of former volcanic activity, contain appreciable quantities of potassium. The water of Owens Lake, in eastern California, yields almost 3 grams of potassium chloride per liter. Both soda and potash are now recovered from the new plant at this locality.

Evaporation in the Quaternary lakes of the Lahontan basin in Nevada and California has at many places resulted in deposits of salt of moderate thickness. Changes in drainage among these basins sometimes resulted in the residual brines, richer in potash, being drawn off into a neighboring depression, and thus it happens, as at Searles Marsh, in San Bernardino County, California, that the salt bed, which here is almost 60 feet in thickness and covers an area of at least 11 square miles, is saturated with a strong brine unusually rich in potassium. In the dissolved salts of Searles Marsh there is almost 7 per cent. of K_2O. A large plant is now under construction for the recovery of potash at this place.

Some small lakes in the "sand hill country" of Nebraska and Colorado contain a remarkably high percentage of potassium which it is thought may have become concentrated from the potash resulting from repeated burnings of the grass in the surrounding country. The Nebraska lakes now (1917) yield about one-half of the potash production of the United States.

The earliest source of potassium was, as is well known, ashes of vegetable matter. Seaweed is especially rich in this metal and also contains iodine. Potash salts are now also recovered from the flue dust of cement plants.

Bromine and Calcium Chloride.—The very soluble salts of bromine and calcium chloride are recovered from residual salt brines in Michigan and West Virginia. The principal production before the war came from the Stassfurt salts of Germany.

CHAPTER XVIII

MINERAL DEPOSITS RESULTING FROM PROCESSES OF ROCK DECAY AND WEATHERING[1]

GENERAL CONDITIONS

The uplifted sedimentary beds, the lavas of the volcanoes, the granular crystalline rocks uncovered by erosion—all these, when exposed at the surface of the earth are subject to a series of changes, the sum total of which is called weathering. The agents are water, air, heat, and vegetable and animal life. Water is essential—without it very little decomposition could take place. Oxygen is also essential, and indeed we often speak of weathering as synonymous with oxidation. Carbon dioxide dissolved in water decomposes the minerals and hastens the process of solution. Change of temperature acts mainly by promoting disintegration, most powerfully by the expansion of water when freezing in cracks and crevices, a force sufficient to break and dislodge heavy rock masses. Vegetable life furnishes carbon dioxide and disintegrates the soil by the vital energy in the roots, and bacterial life changes its composition. Animals burrow in the ground, loosening it and effecting chemical changes.

Weathering differs from alteration and metamorphism in that its ultimate result is the destruction of the rock as a unit; its chemical processes are far more radical and intense than those of the depths. Weathering effaces the texture of the rocks and segregates their chemical compounds in ways wholly different from those of the other processes mentioned. Metals closely associated in the primary rocks part company and seek new associates. Segregations of large masses of single minerals are usually a result of the process. The ordinary silicates and

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[1] N. S. Shaler, The origin and nature of soils, *Twelfth Ann. Rept.*, U. S. Geol. Survey, pt. 1, 1891.

G. P. Merrill, Rocks, rock-weathering, and soils, 1897.

C. R. Van Hise, Metamorphism, *Mon.* 47, U. S. Geol. Survey, 1904.

F. K. Cameron and J. M. Bell, The mineral constituents of the soil solution, *Bull.* 30, Bureau of Soils, Washington, D. C., 1905.

T. L. Lyon, E. O. Fippin and H. O. Buckman, Soils, New York, 1916, p. 764.

the carbonates of iron, magnesium, and calcium are unstable in the belt of weathering. The uppermost thin mantle of the products of weathering we call the soil; in it the disintegration and chemical changes are carried to their limit; it is mixed with the products of life, and its constituents and reactions are, of course, of more interest to the agricultural chemist than to the student of ore deposits.

The depth to which weathering extends differs greatly; in some desert regions, recently glaciated areas, or areas covered by fresh lava flows it is practically absent, disintegration being the only visible effect. In regions of heavy vegetation and rainfall the weathering may extend to a depth of 100 or even 200 feet; along fractures in particularly permeable and soluble rocks like limestone oxidation may be carried to still greater depth; in mineral deposits its effects are in places felt for several hundred or in extreme cases as much as 2,000 feet. As a rule, however, weathering does not extend deeper than 50 feet, and its more intense effects are usually limited to the zone above the surface of underground water.

Disintegration and decomposition work together, but the former is likely to extend deeper than the latter. The upper layers, ordinarily colored red or brown by ferric iron, gradually change into paler-colored, more or less softened and disintegrated rock. In some areas, notably over limestone strata, there is a sharp change to the unaltered rock—so sharp, indeed, that the red clayey soil has often been taken for a different formation resting on the calcareous rock.

Erosional transportation attends disintegration, and removal of material by solution accompanies decomposition, both tending strongly to reduce the volume of the rock. On the other hand, hydration and the peculiar quality of adsorption which the soils possess tend to increase the volume. On the whole weathering lessens the volume. According to G. P. Merrill the granites of the District of Columbia may lose by weathering 13.5 per cent. of their volume; T. L. Watson calculates the loss of granites of Georgia at 7 to 72 per cent. The most marked loss is the shrinking in residual clays derived from limestone; often it is more than 95 per cent. Whitney long ago arrived at the conclusion that 1 meter of residual clay in Wisconsin was derived from a thickness of 35 to 40 meters of limestone or shale.

Except in the easily soluble rocks the decomposition is never complete, for, as brought out by Cameron and Bell, even in the fine soils abundant grains of the original minerals remain unaltered. Other conditions being equal, weathering is most complete in tropical and moist countries. In the United States the most intense action of this kind has taken place in the Appalachian region south of the glaciated area, and this region contains the majority of ore deposits caused by weathering.

Air contains approximately by volume 21 per cent. oxygen, slightly less than 79 per cent. nitrogen (with argon), and 0.03 per cent. carbon dioxide. In the air contained in rain water both oxygen and carbon dioxide are greatly concentrated. In the soils carbon dioxide and air are absorbed; soils and clays of various kinds contain from 14 to 40 cubic centimeters of gas per 100 grams, with 14 to 34 per cent. of carbon dioxide and considerably less oxygen than the air—indeed, in some soils oxygen appears to be absent.[1] Decaying vegetation still further increases the percentage of carbon dioxide. As the ground-water level is approached the oxygen decreases rapidly, as shown by the measurements made by B. Lepsius[2] in bore-holes, and below this level there is probably little left.

Naturally the processes of weathering are hastened by the presence of sulphuric acid derived from the decomposition of pyrite or exhaled from solfataric vents. W. Maxwell[3] has shown interestingly how extensive a part this acid plays in the development of soils on the slopes of volcanoes.

The processes characteristic of weathering are oxidation, hydration, and solution. In the surface waters calcium and magnesium carbonates ordinarily prevail, with a considerable amount of alkaline carbonates and relatively much soluble silica, both derived from the decomposition of the silicates. Under special conditions, as in volcanic regions or in sediments rich in salts, the surface waters may be materially different in composition, being predominantly sulphate solutions. The ground waters contain in addition small amounts of iron and manganese, carried mainly as bicarbonates, also phosphoric acid and sodium chloride.

In the weathered zone will remain the residual, almost insol-

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[1] Cameron and Bell, *op. cit.*, p. 26.

[2] Quoted in F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 477.

[3] W. Maxwell, Lavas and soils of the Hawaiian Islands, Honolulu, 1898.

uble minerals, like quartz, hydrated aluminum silicates more or less closely approaching kaolinite in composition, ferric oxides (as limonite, göthite, or hematite), and manganese dioxide, all mingled to form a red or brown clayey soil.

All these reactions involve the development of colloid bodies like aluminum silicates and hydroxides of iron, which before their transformation into crystalline minerals are characteristic absorbents of many salts. The colloids of manganese, for instance, have a tendency to adsorb potassium and barium. The zone of weathering has indeed been called the realm of the colloids.

DECOMPOSITION OF MINERALS

The silicates of the rocks are decomposed by water rather than dissolved, for the resulting solution does not usually contain the elements in the same proportions as the original mineral. Owing to hydrolysis the solution in most cases gives an alkaline reaction.[1]

Cameron and Bell believe that the more rapid decomposition shown to take place in the presence of CO_2 is in large part due to its combination with the hydrolyzed bases by the formation of bicarbonates rather than to direct solvent action. Orthoclase, for instance, would give $KAlSi_3O_8+HOH=KOH+HAlSi_3O_8$, and the carbon dioxide would form potassium bicarbonate with KOH. Kaolin would form from the unstable silicate $HAlSi_3O_8$.

The biotite, amphibole, and pyroxene, perhaps previously altered to chlorite below the water-level, break up into soluble earthy carbonates, with limonite, hydrous aluminum silicate, and silica, the latter three in colloidal state. These ferromagnesian minerals are attacked first, so that the ordinary surface waters contain more of the carbonates of calcium and magnesium than of other salts. The soda-lime feldspars come next while the alkali feldspars are more resistant. Quartz is only partially attacked. The decomposition of orthoclase is usually expressed in the following equation:

$$2KAlSi_3O_8+2H_2O+CO_2=H_4Al_2Si_2O_9+4SiO_2+K_2CO_3.$$

The ultimate product is kaolin, or allied colloidal bodies. Muscovite or sericite do not result from weathering, although the colloidal aluminum silicate may adsorb potassium and form amorphous compounds related in composition to the white micas.

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[1] F. W. Clarke, *Bull.* 167, U. S. Geol. Survey, 1900, p. 156.

Zeolites are undoubtedly unstable in the zone of weathering. Muscovite or sericite is slowly attacked; Cameron and Bell[1] treated 2 grams of powdered muscovite with a liter of pure water for 14 months in paraffine cylinders and obtained in the solution 10.4 parts K per million; when treating it with carbon dioxide in water they obtained 18.3 parts per million. The same quantity of orthoclase with pure water yielded a solution with 1.7 parts per million of K; when saturated with CO_2 it yielded 2.5 parts per million. Muscovite thus yields its potassium more easily than orthoclase. Albite treated in the same way gave 1.0 and 1.1 parts of sodium per million respectively. Earlier experiments leading to similar results have been undertaken by A. Daubrée, R. Müller, A. S. Cushman and F. Henrich.[2]

Magnetite is soluble with difficulty, but finally yields hematite and limonite. Pyrite easily yields limonite and sulphuric acid. Apatite appears to be rather easily soluble, especially in carbonated water. Cameron and Bell[3] treated powdered chlorine apatite suspended in water at 25° C. for 7 days, passing CO_2 through the liquid. The solution showed 0.0378 gram CaO, 0.0096 gram P_2O_5, and 0.0026 gram hydrochloric acid per liter of solution. In soils and clays the phosphates are decomposed or hydrolyzed, soluble phosphates being formed, but the percolating water contains these only in minimal quantities. It is stated[4] that humus suspended in water can adsorb calcium and a considerable amount of phosphoric acid from the calcium phosphates.

The reactions of the iron phosphates are in the main similar to those of the calcium salts. Lachowicz[5] found that organic matter in soil is a distinct solvent for ferric phosphate. Cameron and Bell ascertained that carbon dioxide greatly aided the solution of ferric phosphate, 5 grams of which, shaken for 5 days with 100 c.c. H_2O and later with 100 c.c. of saturated solutions

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[1] *Op. cit.*, p. 33.

[2] A. Daubrée, Études synthétiques de géologie expérimentale.

R. Müller, *Jahrb.* K. k. geol. Reichsanstalt, vol. 27, 1887.

A. S. Cushman, *Bull.* 92, Bureau Chemistry, U. S. Dept. Agr., 1905.

F. Henrich, Ueber die Einwirkung von Kohlensäurehattigen Wasser auf Gesteine *Zeitschr. prakt. Geol.*, 1910, pp. 84–94.

[3] *Bull.* 41, Bureau of Soils, U. S. Dept. Agr., 1907.

[4] Patten and Waggaman, Absorption by soils, *Bull.* 52, Bureau of Soils, U. S. Dept. Agr., 1908.

[5] Gesteins und Bodenkunde, 1877, p. 329.

of CO_2 and $CaSO_4$, yielded respectively 74, 171, and 118 milligrams of phosphoric acid.

Zircon, pyrope garnet, tourmaline, and similar minerals are almost insoluble.

Quartz also shows great resistance and appears practically insoluble in the zone of weathering, except when exposed to the action of a stronger solution of alkaline carbonates. C. W. Hayes,[1] M. L. Fuller,[2] and C. H. Smyth[3] have observed a marked corrosion of quartz pebbles in conglomerates, but the exact nature of the reaction is uncertain. Cherty and fine-grained quartz is a little more soluble.[4] The theory of the origin of the Lake Superior iron ores, supported by Van Hise and Leith, is based on the supposed solubility of such material. It was formerly thought that certain organic acids had the power of dissolving quartz, but this is now considered very questionable.

TOTAL CHEMICAL CHANGES BY WEATHERING

The studies and analytical work of G. P. Merrill have greatly advanced our knowledge of weathering, and many others have contributed valuable data. A compilation of a number of characteristic gradational analyses is given in F. W. Clarke's "Data of geochemistry"[5] and allows an estimate of the final effects of weathering. The analyses show consistently an apparent increase in alumina and water and decreases in SiO_2, CaO, MgO, K_2O, and Na_2O; in short the composition tends toward that of a ferruginous kaolin, except that in the weathering of acidic rocks residual quartz prohibits the decrease of silica to the amount characteristic of kaolin. Comparing equal volumes we find little actual change in the quantity of alumina, and for purposes of comparison this oxide is assumed to be constant. By recalculating the analyses on this basis, the percentage lost or gained by each constituent may be ascertained. In the analyses quoted, the loss of silica is the largest, varying from 8 to 32 per cent. by weight of the original rock and from 15 to 52 per cent. of the original quantity of silica. The abstraction of silica

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[1] *Bull.* Geol. Soc. Am., vol. 8, 1897, p. 213.

[2] *Jour. Geology,* vol. 10, 1902, p. 815.

[3] *Am. Jour. Sci.,* 4th ser., vol. 19, 1905, p. 277.

[4] G. Lunge and C. Millberg, *Zeitschr. angew. Chemie,* 1897, p. 393.

[5] *Bull.* 616, U. S. Geol. Survey, 1916, pp. 486–490. See also, C. K. Leith and W. J. Mead, Metamorphic geology, New York, 1915, pp. 3–24.

as soluble silicates is, then, the dominant factor in the weathering of silicate rocks.

Compared with the decrease in silica, the losses of bases in silicate rocks are small. Calcium, magnesium, sodium, and potassium are removed, but the loss is ordinarily only partial. The leaching of both potassium and sodium is characteristic and is markedly different from certain processes in the alteration of wall rocks of ore deposits, where sodium is completely removed and potassium enriched.

The analyses quoted show that from one-seventh to one-fifth of the iron oxides are carried away. The water invariably increases.

RESIDUAL CLAY[1]

Occurrence.—The residual clays, as might be expected, are found mainly where decay of rocks has proceeded unchecked for a long time and where the products have not been swept away by strong erosion or by glacial action.

Such clays are found in all parts of the world; in the United States they occur chiefly in the southern Appalachian region. Acidic granular rocks like granite and gneiss—particularly those rich in feldspar and poor in ferromagnesian silicates, like pegmatite dikes—yield the best and purest clays, but even these must often be purified by washing in order to remove residual quartz and limonite. At varying depths, 100 feet at most, these clays gradually change into unaltered rocks.

Uses and Properties.—The ordinary varieties of residual clays are used for brickmaking, the purer for fire-bricks, the finer grades for pottery; for the last use the deposits of the United

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[1] The literature of clays is exceedingly voluminous. For information more detailed than can be given here, consult:

H. Ries, Clays, occurrence, properties, and uses, 1908.

H. Ries, A review of the theories of origin of white residual kaolins, *Trans*. Am. Ceramic Soc., vol. 13, 1911.

I. E. Sproat, Refining and utilization of Georgia kaolins, *Bull.* 128, U. S. Bureau of Mines, 1916.

H. O. Buckman, The chemical and physical processes involved in the formation of residual clay, *Trans*. Am. Ceramic Soc., vol. 13, 1911.

J. H. Watkins, White-burning clays of the southern Appalachian states, *Trans*. Am. Inst. Min. Eng., vol. 51, 1916, pp. 481–501.

A full bibliography of the older literature by H. Rösler is contained in *Neues Jahrb.*, Beil. Bd. 15, 1902, p. 231.

States are insufficient, hence large quantities are imported from England. These purer grades of white burning clays are usually called kaolin, ball clay, paper clay or china clay, and are also used in manufacturing paper and as fillers in paints, putty and crayons. About 200,000 tons of these fine clays are produced in the United States, the average price being $8 per ton. Their composition,[1] before and after washing, is indicated by the following analyses.

ANALYSES OF CRUDE AND WASHED KAOLIN, WEBSTER COUNTY, SOUTH CAROLINA

	Crude	Washed	Kaolinite
SiO_2	62.40	45.78	46.5
Al_2O_3	26.51	36.46	39.5
Fe_2O_3	1.14	0.28	....
FeO	....	1.08	....
CaO	0.57	0.50	....
MgO	0.01	0.04	....
Alkalies	0.98	0.25	....
H_2O	8.80	13.40	14
Moisture	0.25	2.05	....
Total	100.66	99.84	100
Clay substance	66.14	93.24	....

It is sometimes difficult to determine with the microscope the particles of kaolinite ($H_4Al_2Si_2O_9$) in an altered rock, on account of their minute flaky size and low double refraction. It is probable that kaolinite is present in the residual clays, but besides this well-defined mineral there may be other hydrated silicates of alumina separated in colloidal form, as gels, during the weathering. Among these hardened gels there are some highly hydrated forms like halloysite and allophanite which are decomposed by HCl. Others corresponding approximately to kaolinite are only attacked by H_2SO_4. Both classes occur in sediments as well as in residual deposits. Paper clays are mined in Cretaceous beds of South Carolina and Georgia, plastic kaolins or ball clays are obtained in Tertiary beds in

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[1] H. Ries, Economic geology, 1917, pp. 170–186.

Florida and Western Tennessee. Residual clays are mainly mined in North Carolina.

The most important property of clay is plasticity, by means of which it can be kneaded or molded into a desired shape, which it retains when dry. Not all the residual clays are plastic, nor is the pure mineral kaolinite. It is now generally believed that the plasticity of clay depends upon the presence of colloids.[1]

The tensile strength of air-dried clays varies from 15 to 400 pounds or more per square inch, according to Ries. The fusibility varies according to the impurities present. In low grades of clay incipient fusion may occur at about 1,000° C., while in refractory clays, which are low in fluxing impurities, it may not occur until 1,300° or 1,400° C. is reached. The melting-point of kaolin is about 1,800° C.

Origin.—The best residual clays are derived largely from the decomposition of the feldspars as indicated on p. 322 by carbon dioxide. The process is hastened by sulphuric acid, as is attested by the great development of pure kaolin in the upper levels of pyritic mineral deposits.

The decrease in volume by decomposition of orthoclase, if the silica were liberated in soluble form, would be 54.44 per cent. In kaolinization anorthite simply loses its calcium oxide and takes up CO_2 and H_2O. Pure orthoclase loses 43.24 per cent. SiO_2 and all of its potassium; albite loses 45.87 per cent. SiO_2 and all of its sodium.

The origin of kaolin has been the subject of much discussion. About 10 years ago H. Rösler[2] published a long and important paper which gave rise to an animated discussion.[3] Rösler concludes that kaolin is not formed by weathering, but only by pneumatolytic or allied processes by the action of thermal waters. It is impossible to accept these results and they have been vigorously contested by Stremme and Barnitzke;[4] the latter showed that the celebrated deposit at Meissen, in Saxony, where a high grade of chinaware is made, is decidedly a product of

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[1] H. E. Ashley, The colloid matter of clay and its measurement, *Bull.* 388, U. S. Geol. Survey, 1909, p. 65.

[2] *Neues Jahrb.*, Beil. Bd. 15, 1902, p. 231–393.

[3] See H. Rösler, *Zeitschr. prakt. Geol.*, 1908, p. 251; Stremme, *idem*, p. 122.

[4] Ueber das Vorkommen der Porcellanerde bei Meissen und Halle, *Zeitschr. prakt. Geol.*, 1909, pp. 457–472.

weathering, gradually changing in depth into unaltered porphyry and syenite. The chemistry of kaolin is given in full by H. Stremme,[1] and has lately been summarized by Doelter.[2]

It has been shown by W. Lindgren[3] and others that kaolin does not form in deposits that are due to ascending thermal waters, except possibly very close to the surface, where they may mingle with atmospheric waters. The idea that the mineral may form by pneumatolysis, or the action of water or gases liberated at high temperature from igneous magmas, is assuredly untenable; a strongly hydrous mineral, parting with its water at the comparatively low temperatures of 300° to 400° C., could not possibly originate together with such minerals as topaz and tourmaline. The frequent association of kaolin with cassiterite veins—for instance, in Cornwall—has been held by L. v. Buch, A. Daubrée, J. H. Collins,[4] and H. Rösler to indicate a derivation by the action of hydrofluoric acid on feldspars, but as the kaolin deposits, during the metallogenetic epoch, were under the same general conditions of pressure, temperature, and depth as the tin deposits, this view must be abandoned.

Extensive observations in the United States have shown that in mineral deposits kaolin is scarcely ever a primary mineral, but has been derived largely by the action of sulphuric acid on the feldspar minerals of the rocks and on sericite, which is often abundantly developed in ore deposits formed under widely differing physical conditions. In view of this, it seems odd that Rösler expresses astonishment at finding a large amount of muscovite with the kaolin from Cornwall and suggests that the former may be a secondary product. G. Hickling[5] has investigated the china clays of Cornwall and shows that they form essentially a sheet covering the corroded surface of the granite and that they have resulted from the weathering of sericitic granite, the sericite being due to previous alteration by thermal waters.

Kaolin, then, is formed abundantly in the zone of weathering and in smaller amounts for a considerable distance below this zone.

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[1] Die Chemie des Kaolins, *Fortschritte der Min., Krist. u. Petr.* Jena, 1912.

[2] C. Doelter, Handbuch der Mineralchemie, vol. 2, 1914, pp. 31–91; 125–137.

[3] The origin of kaolin, *Econ. Geol.*, vol. 10, 1915, pp. 89–93.

[4] J. H. Collins, *Min. Mag.*, vol. 7, 1886–1887, p. 217.

[5] *Trans.* Inst. Min. Eng. (England), vol. 36, 1908–1909, p. 10.

RESIDUAL IRON ORES (LIMONITE AND HEMATITE)

Origin.—During the processes of weathering only a small part of the iron is carried away in solution; the greater part remains in the rock altered to limonite ($2Fe_2O_3.3H_2O$), to göthite ($Fe_2O_3.H_2O$), or to indefinite colloidal mixtures of various hydroxides of iron; hematite may also be present. In places basic sulphates or phosphates may remain, as well as somewhat indefinite and unstable ferric silicates. Nontronite, $H_4Fe_2Si_2O_9$, the equivalent of kaolin, is said to be present in weathered rocks. In the zone of weathering the iron shows a strong tendency to move outward and segregate in irregular or mammillary masses, separated by clayey material, though much of it, of course, remains intimately mixed with clay. The same is true of manganese, some of which may be associated with the limonite, though when much manganese is present, it also tends to separate by itself.

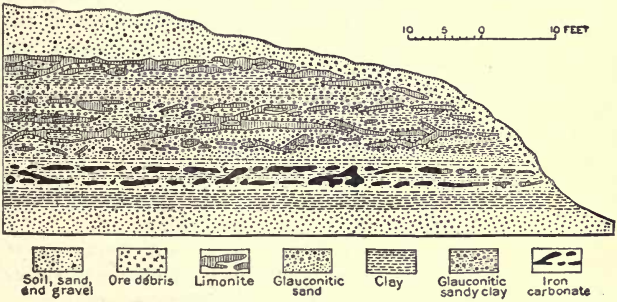

FIG. 104.—Section showing oxidation of iron carbonate to limonite in Tertiary beds, Cass County, Texas. *After E. F. Burchard, U. S. Geol. Survey.*

The "centrifugal" tendency of iron hydroxide is well seen in many oxidized mineral deposits, often also in the weathering of pebbles. A fine instance was observed in the cobbles of andesite in the Tertiary river-bed at Iowa Hill, California. The outside of these cobbles is hard and consists of an impure limonite; the center contains soft yellowish kaolin.

During the concentration the ferric hydroxides (see p. 257) were probably transported as colloids, which hardened and became crystalline, as shown by the radial structure of many concretions. The chemical character of these ores has rarely been studied in detail; probably it will be found that barite, oxidized zinc minerals, and compounds containing manganese, nickel, and cobalt are present. Many of the limonites are rather pure and they are of considerable economic importance.

Classification.—One class of residual brown iron ores is derived from the decomposition of deposits of siderite or pyritic ores, both usually formed by ascending waters, or from the weathering of black bands or glauconite beds (Fig. 104). Such limonites in places reach considerable depths, dependent on the penetrating power of oxygenated waters. The decomposed croppings of pyritic ores are not often used as iron ores.

Another class consists of local segregations of limonite or allied hydroxides in the decayed rock and residual clay near the surface. These masses are particularly common in limestone areas. Little or no siderite is found near the surface, but it may appear in the limestone at greater depth. When oxygen is exhausted the iron is more easily transported as a bicarbonate and the metasomatic replacement of calcite by siderite may then occur. There are, however, few deposits of limonite which change in depth to large irregular replacements of siderite, so that it may be assumed that the rate of solution and downward transportation of the precipitated limonite is slow.

Finally, a third class of residual iron ores, consisting of limonites mixed with hematite, occurs as widespread sheets formed by the gradual decay of strongly ferriferous rocks.

Brown Hematites of the Appalachian Region.[1]—In the United States the residual iron ores are most abundant in the Appalachian region, mainly in Alabama, Georgia, Virginia, and Tennessee. The annual production of such ores is about 2,000,000 long tons, a small part, of course, of the yearly output of iron ores in the United States. These so-called "brown hematites"

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[1] C. W. Hayes and E. C. Eckel, Iron ores of the Cartersville district, Georgia, *Bull.* 213, U. S. Geol. Survey, 1902, pp. 233–242.

E. C. Eckel, Limonite deposits of eastern New York, etc., *Bull.* 260, U. S. Geol. Survey, 1904, pp. 335–342.

R. J. Holden in Mineral resources of Virginia, 1908.

E. C. Harder, The iron ores of the Appalachian region in Virginia, *Bull.* 380, U. S. Geol. Survey, 1908, pp. 215–254.

E. C. Harder and E. F. Burchard, *Mineral Resources*, U. S. Geol. Survey, particularly pt. 2, chapter on Iron, 1908.

E. F. Burchard and E. C. Eckel, Birmingham district, *Bull.* 400, U. S. Geol. Survey, 1910, pp. 145–167.

are mined in many small deposits; their content in iron ranges from 38 to 52 per cent. (limonite 59.89 per cent. Fe); most of them are comparatively rich in phosphorus.

Most of the southern limonites lie in Cambro-Silurian strata and extend along the "Great Valley," between the pre-Cambrian on the east and the Paleozoic rocks on the west. They are classed as valley ores, mountain ores, and Oriskany ores.

The *valley ores* appear as irregular deposits of shallow pockets in clay derived from the decomposition and solution of Cambro-Silurian limestone or dolomite. The ores lie as scattered lumps in the clay, not so much on the eroded surface of the limestone, but rather higher up (Fig. 105). Each deposit is soon exhausted, and few extend below a depth of 50 feet. The ores are mixtures of limonite, göthite, and clay; the composition ranges from 40 to 56 per cent. Fe, 5 to 20 per cent. SiO_2, 0.05 to 0.5 per cent. P, and 0.3 to 2.0 per cent. Mn.

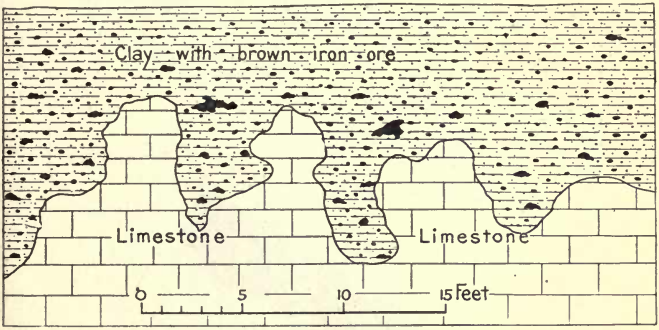

FIG. 105.—Vertical section showing structure of the valley brown ore deposits of the Rich Hill mine, Virginia. *After E. C. Harder, U. S. Geol. Survey.*

Many of these ores were evidently concentrated under conditions different from those of to-day; most of them are probably of Tertiary age as shown particularly in the deposits south of Birmingham, Alabama. It is not unlikely that the same applies to many "mountain ores."

The *mountain ores*, according to Harder, show greater variation in occurrence and appearance. They are found as small discontinuous pockets in residual material above the Lower

Cambrian quartzite at or near the contact with the overlying formation, which is generally a limestone. While these ores are mainly superficial, they are sometimes worked to a depth of several hundred feet. The composition ranges from 35 to 50 per cent. Fe, 10 to 30 per cent. SiO_2, 0.1 to 2 per cent. P, and 0.5 to 10 per cent. Mn. These limonites are often glassy and concretionary.

The occurrences are classed by Harder as follows:

1. Pocket deposits in clay, in part replacements of limestone, in part manganiferous (Fig. 106).
2. Small replacement deposits in shale, along fractures.
3. Deposits in quartzite or sandstone, not abundant, including
   - *a.* Breccia deposits accompanied by replacement.
   - *b.* Vein deposits along faults.

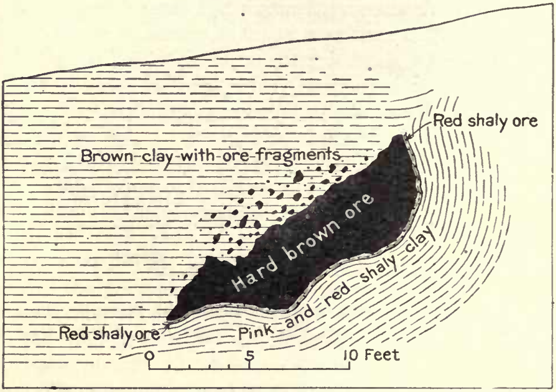

FIG. 106.—Vertical section showing the structure of mountain brown ore occurring as mammillary masses in clay. Mary Creek mine, Virginia. *After E. C. Harder, U. S. Geol. Survey.*

The sandstones of the Cambro-Silurian are often ferruginous in this region. Some of the varieties rich in hematitic cement change along the strike to beds of siliceous hematite, several feet thick and of possible economic importance.

The Oriskany ores[1] are mined in Virginia and form irregular replacements along local folds or fracture zones on the flanks of

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[1] C. M. Weld, The Oriskany iron ores of Virginia, *Econ. Geol.*, vol. 10, 1915, pp. 399–421.

greater anticlines. They occur in the calcareous Oriskany sandstone which is overlain by the Romney shale (*Devonian*) and underlain by the Helderberg limestone (*Silurian*). The ore largely replaces the sandstone, sometimes also the limestone and may be from 10 to 100 feet wide. The greatest depth reached is 600 feet; at this or lesser depth the ore grades into unaltered rock. The iron is considered by some authors to be derived from the Romney shale but is more likely derived from the sandstone itself. The ore is made up of earthy masses and rounded concretions of fibrous limonite filled with clay or sand. The ore from the Callie mine contains about 0.2 per cent. of zinc.[1] At the Callie mine the ore production is about 2,700 tons per month. Probably hematite or turgite are also present for the ore does not contain enough water for limonite; it averages 43 per cent. Fe, 10 to 25 per cent. SiO_2, 0.06 to 0.5 per cent. P, and 0.5 to 4 per cent. manganese. Cobalt and nickel are reported to be present in traces.

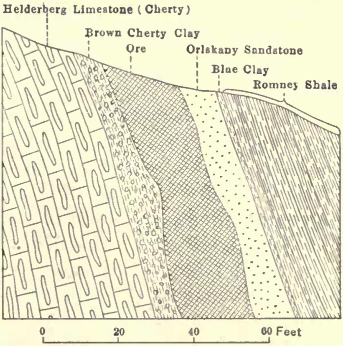

FIG. 107.—Vertical section showing the Oriskany brown ore deposit at the Callie mine, Virginia. *After E. C. Harder, U. S. Geol. Survey.*

The Oriskany ores, like the other "brown hematites," are subjected to a rough concentration in log washers in order to remove the clay.

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[1] The foot-wall limestone is said to contain the same amount of Zn. Letter from S. E. Doak.

Iron Ores of Bilbao, Spain.[1]—The great deposits of Bilbao, in northern Spain, have for many years yielded several million tons annually, the ores being exported to England.

According to Adams, both replacement and residual ores are present. The ores are superficial and limited to areas of Cretaceous limestone, which is 250 feet thick and dips northeast. The white siderite ore, which is found at some depth, is altered near the surface to red hematite with 80 to 90 per cent. Fe_2O_3. The ores are of Bessemer grade. Adams believes that, during the progress of denudation, the calcareous beds became replaced by siderite by the aid of downward-percolating solutions, derived partly from the overlying calcareous shale. Through long-continued rock decay the siderite was altered to hematite and limonite, which now, with much clay, cover the limestone areas like a sheet. One of the largest iron-bearing areas is 2 miles long and 3,300 feet wide; the iron ore in this area had a thickness of about 100 feet.[2]

Residual Ores of Cuba.[3]—Iron ores have been mined for a number of years in the vicinity of Santiago, Cuba, but these ores, of contact-metamorphic origin, consist of hematite with some magnetite and contain a high percentage of sulphur. The three new districts described by Spencer and others are likewise in the eastern part of the island, but are of an entirely different type. They are the Mayari and Moa districts in Oriente province, and the San Felipe in Camaguey (Fig. 108). The ores occur as

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[1] F. D. Adams, Notes on the iron deposits of Bilbao, *Jour.*, Canadian Min. Inst., 1901.

John, *Zeitschr. prakt. Geol.*, 1911, pp. 208–212.

P. Grosch, Geol. Rundschau, vol. 5, 1914–15, pp. 392–400.

[2] Stelzner and Bergeat, Die Erzlagerstätten, vol. 2, 1906, p. 1049, with list of literature.

[3] A. C. Spencer, Three deposits of iron ore in Cuba, *Bull.* 340, U. S. Geol. Survey, 1907, pp. 318–329.

C. M. Weld, The residual iron ores of Cuba, *Trans.*, Am. Inst. Min. Eng., vol. 40, 1909, pp. 299–312.

J. F. Kemp, The iron resources of the world, Int. Geol. Congress, Stockholm, 1910, pp. 793–795; *Trans.*, Am. Inst. Min. Eng., vol. 51, 1916, pp. 3–30.

See also seven papers on the same subject by J. S. Cox, Jr., C. K. Leith, W. J. Mead, A. C. Spencer, C. W. Hayes, W. L. Cumings, B. L. Miller, D. E. Woodbridge, and J. E. Little, in *Trans.*, Am. Inst. Min. Eng., vol. 42, 1911, pp. 73–152. Also C. K. Leith and W. J. Mead, *idem*, vol. 53, 1916, pp. 75–78.

residual mantles resulting from the weathering of serpentine and for the most part lie on plateaus at rather high elevations. They were probably formed during the Tertiary before the uplift of the present plateaus. Near the surface the material is earthy and dark red, sometimes cemented with shot-like lumps of hematite scattered over the surface; underneath lie yellowish ores changing rather abruptly into decomposed and soft serpentine. In places a layer of cherty material is found immediately above the serpentine. In the Mayari district the average depth of the ore is about 15 feet and it extends over an area of 10 by 4 miles. Hundreds of millions of tons are said to be available, allowing for parts of the area which are below the workable grade. The ore is removed by drag-line steam shovels.

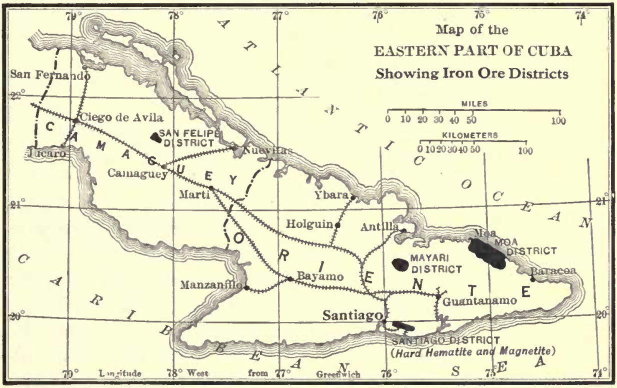

FIG. 108.—Sketch map of eastern part of Cuba. *After W. L. Cumings and B. L. Miller.*

According to analyses the ore is fairly uniform, the metallic iron varying in percentage from 40 to 50. It is remarkably free from phosphorus and evidently contains hematite, limonite, a little magnetite, and also some free aluminum hydroxide. It is, in brief, a typical iron-rich laterite (see p. 351). There is much water; according to Kemp the Moa ores yield 25 to 30 per cent. hygroscopic and 10 to 12 per cent. combined water; silica is low and alumina high. The concentration of nickel and chromium is also remarkable; the latter metal is removed during the smelting; the former is favorable to the quality of the iron.

ANALYSES OF SERPENTINE AND ORE FROM THE MAYARI DISTRICT, CUBA
(*After C. K. Leith*)

	1	2
SiO_2	39.80	2.90
Al_2O_3	1.39	10.24
Fe_2O_3	10.14	72.35
Fe	7.10	50.56
MgO	33.69	......
Cr	0.20	1.66
Ni+Co	0.97	0.84
P	0.001	0.016
S	0.06	0.20
H_2O +	13.31	10.96
	99.561	99.166

1. Serpentine, at depth of 29 feet.
2. Iron ore, at depth of 6 feet.

Analyses by Spanish-American Iron Co.

The porosity of the ore is exceedingly great amounting to 75 per cent. of its volume but lessens near the surface.

In considering the alteration of serpentine to ore in terms of weight it is found that the alumina has remained nearly constant. The changes in the composition of the serpentine during its alteration to ore is shown by Leith and Mead in Fig. 109, which is based on many analyses at uniform intervals. The diagram illustrates the rapid destruction of the serpentine by leaching of SiO_2 and MgO, the marked relative increase of iron and alumina and a gradual loss of nickel. Toward the surface hematite (with a little magnetite) develops from limonite and bauxite from kaolin. In the middle part of the ore body iron has increased in proportion to the alumina, owing probably to re-deposition and oxidation of ferrous iron dissolved by the reducing action of the vegetation. Silica is lost throughout and magnesia is wholly removed.

In 100 pounds of typical serpentine there are 1.5 pounds of alumina and 10 pounds of ferrous oxide. When the magnesia and silica are removed in solution and the iron oxidized there remain approximately 11.75 pounds of limonite, 3.8 pounds of bauxite and kaolin, and, at the most, 2 pounds of minor constituents. This residual of 17.55 pounds contains 7.8 pounds, or 44.4 per cent., of metallic iron and is an iron ore.

Distribution and Stability of Residual Iron Ore.—The residual iron ores are widely distributed in countries of warm climate, where secular decay has progressed without interruption for a long time. It seems, however, that great concentration has been effected only from relatively soluble rocks like limestone and serpentine. Many of the laterites of India, Africa, and other tropical countries are rich in ferric oxide and have the same characteristic concretionary pellets and shots on the surface. Extensive limonite deposits similar to those of Cuba have lately been discovered in Borneo and on Mindanao, in the Philippines.

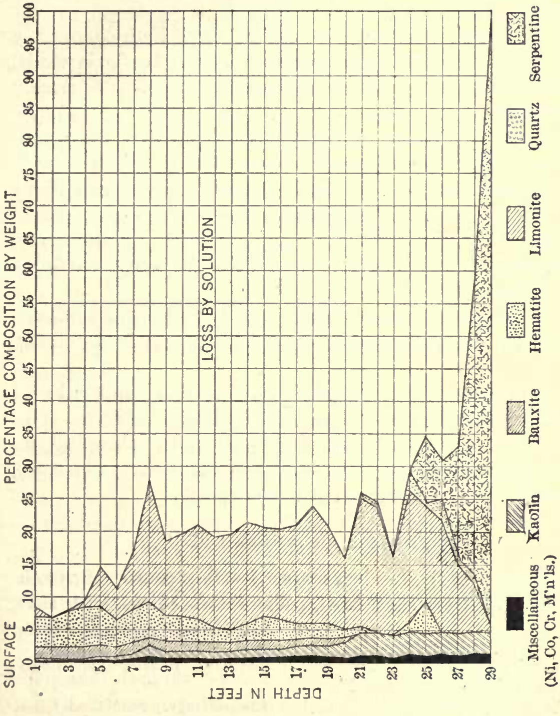

FIG. 109.—Diagram representing in terms of weight the mineralogical changes in the alteration of serpentine rock to iron ore at Mayari, Cuba, on the assumption that alumina has remained constant. *After C. K. Leith and W. J. Mead.*

Vegetation plays an important part in the origin of many of these deposits. Underneath the mat of roots and decayed vegetation the soil in tropical countries is often white or yellowish, indicating that the iron is in the ferrous state, probably as carbonate. When, as happened on the high volcanic plateau of Molokai, Hawaiian Islands,[1] the vegetation is destroyed the soil immediately turns red and hard and shows characteristic pellets of ferric oxide. In part at least the rock is thus changed directly to hematite without passing through the intermediate stage of limonite.

According to H. Wölbling,[2] the natural ferric hydroxides have great stability and cannot readily be changed to ferric oxide, probably not by exposure to air and salt solutions. The freshly precipitated hydroxides are, however, easily converted to ferric oxide and these colloids may easily be crystallized.[3] His experiments show that by the precipitation of ferric solutions with calcite or siderite at 100° C., Fe_2O_3 is easily formed, containing only 1 or 2 per cent. H_2O, while during slow and wet oxidation of ferrosalts, ferric hydrates of iron are obtained. Wölbling also asserts that there are yellow forms of Fe_2O_3, as well as red forms of the hydroxides.

It is certain, at any rate, that the ferric oxide, as well as the hydrates, is very stable when once formed and is not easily altered.

No one can fail to be impressed by certain similarities of the Cuban residual ores to those of the Mesabi range (p. 366). Similar large expanses of rock, weathered under a tropical sun and covered by residual ferric oxide, undoubtedly yielded the material for the sedimentary hematite deposits.

RESIDUAL MANGANESE ORES[4]

The minerals of the residual manganese ores consist of pyrolusite (MnO_2, 63.2 per cent. Mn), psilomelane (MnO_2, with H_2O, K_2O, and BaO; 49 to 62 per cent. Mn), wad (perhaps MnO_2.

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[1] W. Lindgren, The water resources of Molokai, *Water-Supply Paper*, 77, U. S. Geol. Survey, 1903, p. 19.

[2] H. Wölbling, Bildung der oxydischen Eisenerzlager, *Stahl und Eisen*, 1909, p. 1248; also *Zeitschr. prakt. Geol.*, vol. 17, 1909, p. 495.

[3] C. Doelter, Ueber Umwandlung amorpher Mineralkörper in Krystalline, *Tsch. M. und p. Mitt.*, vol. 28, 1909, pp. 556–559.

[4] R. A. F. Penrose, Jr., Manganese, its uses, ores, and deposits, *Ann. Rept.* Arkansas Geol. Survey, vol. 1, 1890.

T. L. Watson, *Trans.*, Am. Inst. Min. Eng., vol. 34, 1904, p. 207.

T. L. Watson, Preliminary report on the manganese deposits of Georgia, *Bull.* 14, Georgia Geol. Survey, 1908.

nMnO+H_2O, varying percentage of metal), more rarely braunite ($3Mn_2O_3.MnSiO_3$ (?); 69.7 per cent. Mn), and manganite ($Mn_2O_3.H_2O$, 62.4 per cent. Mn).

The most common ores are pyrolusite and psilomelane, both occurring frequently in botryoidal, reniform, or mammillary concretions. Harder has shown that these two minerals may form alternating layers in the concretions. Earthy or rough, slaggy forms are also common. Like limonite they are largely colloid deposits, later converted into crystalline minerals.

Primary Sources.—Nearly all workable manganese deposits are of secondary formation—that is, they are concentrated from manganese minerals more sparsely distributed in rocks. Pyrolusite, psilomelane, and wad are always secondary, formed under the influence of weathering, even where they descend to considerable depths below the water level.

In igneous rocks manganese is always present but only in small amounts. The largest percentages (about 0.36 per cent.) are found in syenite and its porphyries and in basalts.

Sedimentary rocks may contain manganese in the form of oxide and carbonate. Manganese nodules occur in some deep-sea deposits.

Analyses of limestones often show a small amount of manganese. In many cherts and jaspers of the sedimentary series manganese is characteristically present as rhodonite or rhodochrosite. On previous pages it has been shown that important deposits of manganese may be produced by sedimentation.

In crystalline schists, especially in those of more basic composition, small quantities of manganese are found.

In some crystalline schists spessartite (manganese garnet), rhodonite, and piedmonite (manganese epidote) appear in considerable quantities.

Finally, rhodochrosite and rhodonite are rather common in ore deposits of hydrothermal or contact-metamorphic origin, and much manganese is present in some metamorphic specularite and magnetite deposits.

E. C. Harder, Manganese deposits of the United States, *Bull.* 427, U. S. Geol. Survey, 1910 (with bibliography and notes on foreign occurrences).

E. C. Harder and D. F. Hewett, *Mineral Resources*, U. S. Geol. Survey, annual publication.

D. F. Hewett, Some manganese ore in Virginia and Tennessee, *Bull.* 640, U. S. Geol. Survey, 1916, pp. 37–71.

Manganese Deposits in the United States.—From the rocks above mentioned manganese may be concentrated by processes of weathering, and its ores are found in concretions embedded in residual clay or ocher and accompanied more or less closely by limonites. During this process some other metals, notably nickel, cobalt, zinc, and barium, have a tendency to accompany the pyrolusite and psilomelane. In general such deposits are superficial or of slight depth and closely parallel the residual limonites already described.

In California small deposits of secondary manganese ores occur in areas of the radiolarian cherts or jaspers of the Franciscan formation (Jurassic ?)

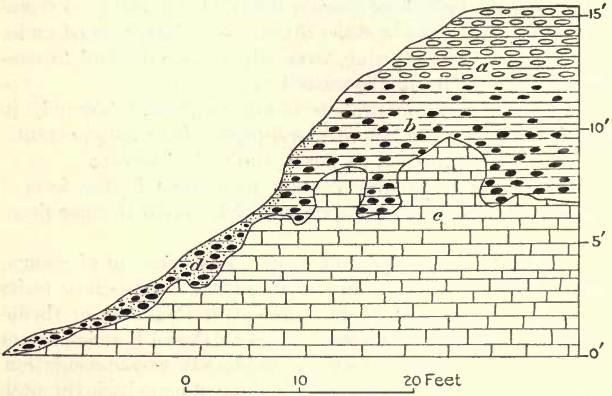

FIG. 110.—Generalized section showing the occurrence of manganese ore at Batesville, Arkansas. *a*, Boone chert (Mississippian); *b*, Cason shale with manganese deposits (Ordovician); *c*, Polk Bayou limestone (Ordovician); *d*, surface clay with manganese deposits. *After E. C. Harder, U. S. Geol. Survey.*

In Arkansas residual ores have been mined at Batesville,[1] where they occur both in the Cason manganiferous shale, of upper Ordovician age, and in clay derived from this formation (Fig. 110). Penrose believed that the manganese was derived from the pre-Cambrian area in southeast Missouri and de-

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[1] R. A. F. Penrose, Jr., *op. cit.*

posited in the sedimentary formation, but the later work of Ulrich and others has shown that erosional epochs have intervened within the formation period assumed by Penrose and that the ores are original marine deposits, reconcentrated during two subsequent land stages, first during the late Silurian and Devonian partial emergence, and second during the post-Paleozoic erosion of the Boone chert.[1]

In the Appalachian region small deposits occur in granites and schists of the Piedmont region, but chiefly in the Paleozoic sediments of the Cambro-Silurian belt—that is, in the general area of the residual iron ores. At the Crimora deposit, in Virginia (Fig. 111), the ore is found as "masses of various sizes scattered through variegated clays in an elliptical basin in a canoe-shaped syncline of the Cambrian quartzite," into which the manganese penetrates as dendritic forms and crystalline coatings.[2]

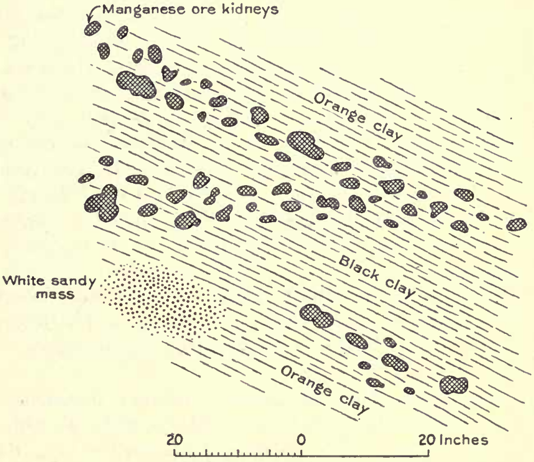

FIG. 111.—Sketch showing distribution of manganese ore lumps in clay at the Crimora mine, Virginia. *After E. C. Harder, U. S. Geol. Survey.*

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[1] E. C. Harder, *Bull.* 427, U. S. Geol. Survey, 1910, p. 117.

[2] E. C. Harder, *idem*, p. 60.

ANALYSIS OF BEST QUALITY CRIMORA ORE

[T. L. Watson, Mineral resources of Virginia, p. 248]

MnO_2	81.703	BaO	0.829
MnO	7.281	CaO	0.880
Fe_2O_3	0.533	MgO	0.630
CoO	0.354	P_2O_5	0.171
NiO	0.096	$(NaK)_2O$	0.467
ZnO	0.623	H_2O[1]	3.405
Al_2O_3	0.896	SiO_2	2.132
		Total	100.000
		Mn	57.297

The manganese deposits of the Appalachian region occur in a decomposed surface zone of many different rocks (Figs. 112 and 113), but most of the deposits are, according to Harder, associated with the top stratum of an impervious Cambrian quartzite overlain by limestone. Penrose[2] holds that they were laid down in local basins during the deposition of the rocks in whose residual clays they are now found. Harder[3] believes that the metal was in the first place obtained from the crystalline rocks of the Piedmont region and that since the emergence of the sediments repeated concentration by rock decay has been going on.

In central Texas,[4] in Mason, Llano, and San Saba counties, oxidized manganese ores occur as products of weathering of crystalline schists containing spessartite, piedmontite, and tephroite.

As stated above, many ore deposits contain manganese as carbonate and silicate, and in the oxidized zone the metal is often highly concentrated in the form of psilomelane, etc., mixed with limonite; these ores often contain gold and silver, but rarely much copper, lead, or zinc. Considerable quantities of such ores, used in part as flux for lead smelting and in part, if of high grade, for the manufacture of spiegeleisen, are mined at Leadville, Colorado.[5] Here the oxidized ore is apparently derived from a manganiferous siderite.

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[1] Probably by difference.

[2] R. A. F. Penrose, Jr., *op. cit.*

[3] E. C. Harder, *op. cit.*, pp. 99–101.

[4] R. A. F. Penrose, Jr., *op. cit.*, p. 432; Sidney Paige, *Bull.* 450, U. S. Geol. Survey, 1911.

[5] S. F. Emmons and J. D. Irving, *Bull.* 320, U. S. Geol. Survey, 1907, p. 26.

The largest part of the manganese obtained in the United States is derived from ores of the Lake Superior region, where manganese occurs as oxides associated with specularite, and from the zinc residues of the great zinc deposit of Franklin Furnace, New Jersey, where the manganese is contained in the franklinite [$(Fe,Zn,Mn)O.(Fe,Mn)_2O_3$] associated with zincite [$(Zn,Mn)O$] in a deposit of deep-seated, probably contact-metamorphic, origin.

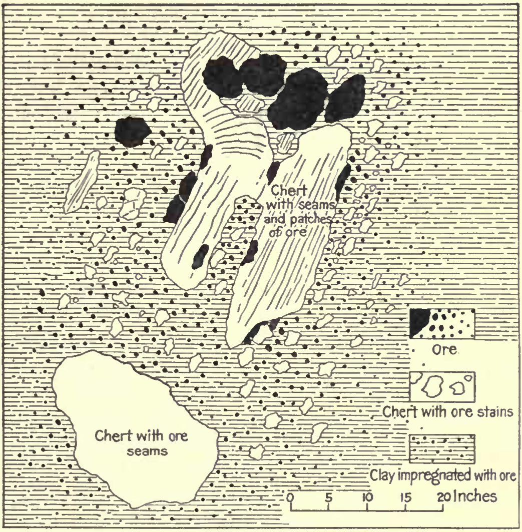

FIG. 112.—Sketch showing occurrence of manganese breccia ore at Reynolds Mountain, Virginia. *After E. C. Harder, U. S. Geol. Survey.*

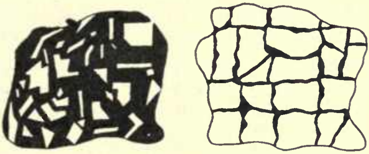

FIG. 113.—Sketch showing development of breccia ore by replacement. White areas, chert or sandstone; black, manganese ore. One-fifth natural size. *After T. L. Watson.*

Brazil.—The high-grade manganese deposits of Minas Geræs, Brazil, have been described by J. C. Branner and O. A. Derby.[1] In the main they appear to be residual ores derived from the weathering of lenses in the crystalline schists containing rhodochrosite, tephroite, and spessartite. The ores are concretions, masses, and vein-like deposits of psilomelane in the soft decomposed rock.

India.—Manganese ores are extensively distributed in India and their occurrence and origin have recently been described in a detailed manner by L. L. Fermor.[2] To a large extent these rich ores are formed by the combined replacement and decomposition of Archean rocks containing manganese silicates. In part the rocks are crystalline schists with spessartite and rhodonite, in part probably non-metamorphosed peculiar igneous rocks, one of which, for instance, consists of spessartite (spandite) and orthoclase with 3.70 per cent. apatite. To a smaller extent the ores are contained in jaspery quartzites and also in laterite, which is purely residual.

Many deposits of the first class contain enormous masses of psilomelane, pyrolusite, and braunite; during the process of weathering almost all the silica and alumina have been removed. Fermor finds no evidence that the alteration has been caused by sulphuric acid, but holds that in some manner, not yet fully understood, it has been effected by surface waters.

Many of the deposits extend to depths far below the water level and Fermor believes that the oxidation may be of very ancient date, perhaps Archean. In some ways these concentrations by surface waters recall the Lake Superior iron deposits.

Origin.—The manganese ores here described as products of weathering and rock decay are in the main similar in origin to the corresponding deposits of iron ore. It is explained on page 272 that iron and manganese, although acting in a similar manner, are usually laid down separately in residual and sedimentary deposits because of the greater solubility of the manganese carbonate. Where sulphates are present the ferrous salt is decomposed easily by oxygen, while manganese sulphate requires

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[1] Literature summarized by E. C. Harder, *Bull.* 427, U. S. Geol. Survey, 1910, p. 183.

[2] L. L. Fermor, The manganese ore deposits of India, *Mem.*, Geol. Survey India, vol. 37, 1909.

the presence of calcium carbonate or some such mineral.[1] On the whole manganese is not transported far from its original source and is characterized by a strong tendency to segregation into local concretions and masses. It is believed that in the main the ordinary surface waters effected the concentration and that the metal has been transformed through the intermediate stage of carbonate.

Production and Uses.—The normal domestic output of manganese ores containing above 35 per cent. Mn was small. Forced production under war conditions has increased the output to about 300,000 tons (in 1918) which is one-third of the amount normally needed. Heavy imports come from Brazil and India.

For the manufacture of spiegeleisen, an alloy with iron containing less than 20 per cent. Mn, low grades of manganiferous iron ore may be used, but for other purposes the ores should contain at least 40 per cent. Mn, less than 12 per cent. SiO_2, and less than 0.3 per cent. phosphorus.

The higher grades of manganese ores are used extensively for the manufacture of ferromanganese alloys, which are employed for many purposes in the iron-smelting industry, particularly for hardening steel.[2] The pure manganese dioxide ores also find an extensive chemical use, for the generation of chlorine and for the manufacture of cells for dry electric batteries.

RESIDUAL BARITE

Barite as residual material and nodular concretions is not uncommon in the residual soils of Virginia and Georgia and in Washington County, Missouri. In Virginia the Cambro-Silurian limestone, according to T. L. Watson, generally contains a notable percentage of barium, and in many places in Georgia the Weisner sandstone, of the same age, also carries barium suphate. In Missouri the barite is concentrated in the soil from veins in the Ordovician Gasconade limestone. The barium may have been transported as the carbonate, which is slightly more soluble than the sulphate, and precipitated by water carrying sulphate. Much of the barite produced in the United States is obtained from residual clays (p. 376).

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[1] F. P. Dunnington, *Am. Jour. Sci.*, 3d ser., vol. 36, 1888, p. 177.

[2] *Mineral Resources*, U. S. Geol. Survey, pt. 1, 1908, p. 138, and in later issues.

RESIDUAL ZINC ORE

In the Appalachian region, in western Virginia and eastern Tennessee, the Cambro-Silurian limestones contain in places sulphides of lead and zinc distributed in brecciated and crushed zones. At such localities the deep residual soil often contains calamine and smithsonite, the hydrated silicate and the carbonate of zinc, with some cerussite and galena. These ores occur next to the limestone at the bottom of the clay (Fig. 114), not scattered through it like limonite and pyrolusite.[1]

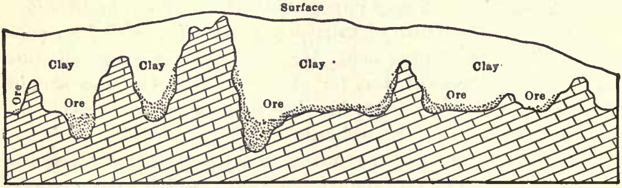

FIG. 114.—Section in open cut at the Bertha zinc mines, Virginia, showing relations of the residual ore to the limestone chimneys and the residual clay. *After T. L. Watson.*

RESIDUAL OCHERS[2]

The residual ochers are impure deep-red, yellow, or brown pulverulent materials consisting usually of predominant limonite and hematite with more or less clay and are generally used for pigments. They are no doubt colloid precipitations. The terms Indian red, sienna, and umber, the latter two for the darker yellowish-brown and brown shades, are in use. Not all mineral pigments are natural products, for roasted pyrite, siderite, slates,

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[1] W. H. Case, The Bertha zinc mines at Bertha, Virginia, *Trans.*, Am. Inst. Min. Eng., vol. 22, 1894, pp. 511–536.

T. L. Watson, Lead and zinc deposits of Virginia, *Bull.* 1, Virginia Geol. Survey, 1905.

T. L. Watson, Mineral resources of Virginia, 1907.

T. L. Watson, Lead and zinc deposits of the Virginia-Tennessee region, *Trans.*, Am. Inst. Min. Eng., vol. 36, 1906, pp. 681–737.

[2] E. F. Burchard and J. M. Hill, *Mineral Resources*, U. S. Geol. Survey, annual publication, "Mineral Paints."

G. P. Merrill, Non-metallic minerals, 1910, pp. 104–111.

and shales are also used.[1] The southern Clinton iron ores are also employed for these purposes.

The residual iron ore deposits of the Southern States contain material which may be classed and is used as ocher. Especially interesting are the Cartersville deposits,[2] in Georgia. These ochers occur only in the Weisner (Cambro-Silurian) quartzite, in the lower part of the residual zone immediately above the yet solid rock, and also in shattered zones in the quartzite itself. The quartzite contains about 90 per cent. SiO_2, 1.5 per cent. FeS_2, 0.5 per cent. Fe_2O_3, and also an unusual percentage of barium sulphate (4.46 per cent. in the analysis given by Watson). The calculated constituents of the ocher are 66 per cent. limonite, 25 per cent. clay, and 9 per cent. quartz; a little hematite is probably also present.

Hayes and Watson are in agreement regarding the origin of the ocher, considering it as resulting from a metasomatic replacement of the cement and the quartz grains of the quartzite by limonite. The process begins by the permeation of the grains by dendritic limonite. This direct formation of the ocher is scarcely probable, but more likely it has progressed by means of an intermediate stage of siderite. The replacement of quartz by iron carbonate is a well-known phenomenon, illustrated, for instance, in the Coeur d'Alene lead deposits of Idaho.

The annual domestic production of natural pigments amounts to about 57,000 tons. The mining is done mainly in open pits, and the material is crushed, washed in a log-washer, and allowed to settle in tanks.

RESIDUAL PHOSPHATES

As described more fully on page 275, many sedimentary beds contain much phosphate of calcium, often in oölitic or concretionary form. When these beds are exposed to surface waters an enrichment usually takes place by solution of calcium carbonate, provided the beds are permeable to the circulating

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[1] B. L. Miller, The mineral pigments of Pennsylvania, *Rept.* No. 4, Topographic and Geologic Survey Commission of Pennsylvania, Harrisburg, 1911.

F. T. Agthe and J. L. Dynan, Paint-ore deposits near Lehigh Gap, Pennsylvania, *Bull.* 430, U. S. Geol. Survey, 1909, pp. 440–454.

[2] C. W. Hayes, Iron ores in the Cartersville district, Georgia, *Trans.*, Am. Inst. Min. Eng., vol. 30, 1901, pp. 403–419.

T. L. Watson, The ocher deposits of Georgia, *Bull.* 13, Georgia Geol. Survey, 1906.

waters. Many important phosphate deposits—for instance, those of Florida, South Carolina, and Tennessee—have been thus enriched.

DEPOSITS OF HYDRATED SILICATES OF NICKEL

The original home of nickel, cobalt, and chromium is in the peridotitic and pyroxenic rocks and in the serpentines derived from them, although traces of these metals are also frequently noted in analyses of other basic rocks. The primary condition of the nickel in the rocks is not always known; probably it occurs both as silicate and as sulphide, the latter in microscopic grains, the former as an admixture in iron-magnesium silicates. From the serpentines and peridotites the nickel is sometimes concentrated in commercially important quantities by processes of weathering and the ores thus formed are always the green hydrated silicates of nickel. Chromite, which always occurs in these basic rocks, does not readily yield oxidized minerals in the zone of weathering. Sulphates of chromium have been observed in a quicksilver mine in California, but no silicate analogous to garnierite exists.

Nickel silicates are diverse and uncertain in composition. The most important are genthite, $H_4Ni_2Mg_2(SiO_4)_3.4H_2O$; connarite, $H_4Ni_2Si_3O_{10}$; and garnierite, $(Mg,Ni)SiO_3+nH_2O$. According to an analysis by A. Liversidge garnierite contains 38.35 per cent. SiO_2; 32.52 per cent. NiO; 10.61 per cent. MgO; 0.55 per cent. Al_2O_3 and Fe_2O_3; 11.53 per cent. H_2O (at red heat) and 6.44 per cent. H_2O (at 100° C.).

Such deposits are superficial and the oxidizing surface waters have been the carrying and concentrating agency. The ores rarely extend far below the water-level and in some cases are contained in the residual clays of the completely weathered rock. These nickel ores are often accompanied by cobalt in the form of separate masses of asbolite, a rather indefinite mixture of hydrous oxides of manganese and cobalt.

These deposits do not contain sulphides, and copper is rarely present. The accompanying minerals are quartz, chalcedony, opal, and various obscure hydrous magnesium silicates, sometimes also a little magnesite. Nickel ores of this kind are not uncommon, but have attained commercial importance only in New Caledonia.

The nickel mine at Riddles, in southern Oregon, has been described by several authors.[1] The parent rock is a peridotite containing 0.10 per cent. of NiO. The olivine separated from the rock contained 0.26 per cent. of NiO and all observers agree that the nickel ores are formed from this silicate. In the finest joints of the rock silica and nickel-magnesium silicates are deposited, and between them lies the oxidized rock converted to a limonite with some clay and chromite.

One of the two most important nickel-bearing districts of the world is in New Caledonia.[2] The island is about 250 miles long and 30 miles wide; one-third of the area is underlain by post-Cretaceous serpentine and peridotite. The lower slopes are covered by a deep mantle of decayed rock ("variegated clay") which really is an iron ore containing, in per cent., 18 silica, 69 ferric oxide, 0.45 alumina, 1.64 nickel oxide and 10 water. The garnierite deposits are found at elevations of from 400 to 2,500 feet, sometimes on fairly steep slopes, or in the saddles of ridges and spurs. Underneath the "variegated clay" at depths of from 20 to 75 feet the nickel ores occur often descending into the serpentine along fissures and accompanied by chalcedony and opal. There are many small deposits; the largest contained only 600,000 tons. The ores are worked by open cuts and carefully graded and sorted. Glasser classifies the deposits in vein-like, brecciated, impregnations and earthy masses. In the latter there is much dark brown "chocolate ore" in which the green silicate is not visible. The clayey ore averages, in per cent., 23 water, 5–7 nickel oxide, 10–12 ferric oxide, 25 magnesia, 40 silica, no lime, 1.1 chromic oxide, 0.12 cobalt, and 1.5 alumina. Most of the ore is exported. Some of it is sun dried and briquetted for local smelting to 45 per cent. nickel matte with limestone and gypsum flux.

The New Caledonia deposits were discovered by the geologist Garnier in 1864; the mines were opened 10 years later, and the cheaply mined rich ores made all nickel deposits elsewhere unprofitable. In 1906 the maximum output of 144,000 metric tons

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[1] Diller and Clarke, *Bull.* 60, U. S. Geol. Survey, 1890, p. 21.

G. F. Kay, *Bull.* 315, U. S. Geol. Survey, 1907, p. 120.

[2] E. Glasser, Rapport sur les richesses minérales de la Nouvelle Calédonie, *Ann. des Mines* (10), vol. 5, 1904, pp. 29–154, 503–701.

G. M. Colvocoresses, *Eng. and Min. Jour.*, Sept. 21 and 28, 1907.

W. G. Miller, Nickel Deposits of the World, reprinted from Report of Royal Ontario Nickel Commission, Toronto, 1917, pp. 234–264.

was reached. Lately, owing to the active competition of the Sudbury mines (p. 814), the output has been materially reduced. In 1916, 30,100 tons of ore as well as 5,000 tons of nickel matte were exported to England and France. A small quantity of cobalt ore, a black, earthy asbolite, was exported for a number of years but at present can not compete with the ore from Cobalt, Ontario. The island also produces much chromite (p. 794).

BAUXITE[1]

Introduction.—Clay, as more or less impure kaolin, is the most abundant product of rock decay, but although it carries 39.8 per cent. alumina its use as a source of metallic aluminum has not been found possible. Corundum is not abundant enough to be used for this purpose. Cryolite (Na_3AlFl_6), a mineral obtained from pegmatitic masses occurring in Greenland, was formerly an important aluminum ore and is still used, in smaller quantities, in the electrolytic processes for the extraction of aluminum.

In certain places the weathered zone, however, contains the hydroxides of aluminum and of these bauxite is the most important aluminum ore. There are three aluminum hydroxides: Diaspore, $Al_2O_3.H_2O$, with 85 per cent. Al_2O_3; gibbsite or hydrargillite, $Al_2O_3.3H_2O$, with 65.4 per cent. Al_2O_3; and bauxite, $Al_2O_3.2H_2O$, with 74 per cent. Al_2O_3.

The independence of bauxite as a mineral species is, however, questioned and many authors consider it a hardened and in part crystallized hydrogel of indefinite composition. The Georgia bauxite, according to T. L. Watson, corresponds well to gibbsite. F. Laur states that the French bauxites also agrees with the formula given above for this mineral. Bauxite forms compact, earthy, also very commonly pisolitic masses, the individual concretions often having a diameter of several centimeters. It is gray, cream-colored, yellowish or brown and

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[1] C. W. Hayes, Bauxite, its occurrence, geology, etc., *Sixteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1895, pp. 547–597.

T. L. Watson, *Bull.* 11, Georgia Geol. Survey, 1904 (Bibliography).

F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, pp. 493–501.

W. C. Phalen, *Mineral Resources*, U. S. Geol. Survey, Annual publication.

J. W. Richards, *Mineral Industry*, Annual publication.

is usually admixed with silica and ferric oxide. Its occurrence and structure lend probability to the view that it has originated as a colloid precipitate. The bauxites contain in places crystalline gibbsite as crusts or veinlets; diaspore has been identified more rarely and quite naturally as it usually formed at higher temperature than that prevailing in residual deposits.

The bauxites always contain titanium, averaging as much as 4 per cent. TiO_2, and some vanadium but in this they merely share the peculiarities of residual and sedimentary clays. Some investigators state that bauxite contains residual rutile while others have failed to find any titanium mineral. Most probably the titanic dioxide is present in colloid state.

Little or no hydroxide of aluminum forms in ordinary rock weathering. Cameron and Bell[1] state that during an examination of several thousand soils from all parts of the United States, hydroxide of aluminum was observed in only one sample, which came from southern California. Bauxite, it may be concluded, is thus rarely formed in the temperate region.

In tropical countries, on the other hand, the deep residual soil very often contains aluminum hydroxide. This has been called laterite (later, brick) and is variously defined. We may say that true laterite is essentially a mixture of the hydroxides of iron and aluminum with more or less free silica, but there are all gradations toward an ordinary ferruginous clay. The laterite may be derived from any igneous or sedimentary rock but serpentine and limestone are specially favorable. The iron ore from Mayari, Cuba (p. 334) is a laterite exceptionally rich in iron. Many so-called laterites are not true residual but transported deposits. Laterites may or may not contain bauxite of economic value; they have been described from many lands and the literature is very extensive.[2]

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[1] *Bull.* 30, Bureau of Soils, U. S. Dept. of Agr., 1905, p. 28.

[2] A. Streng, Zeitschr. deutsch. Geol. Gesell., vol. 39, 1887, p. 621. (Germany).

A. Bauer, Neues Jahrb., Festband, 1907, p. 33 and 1898, pt. 2, p. 192. (Seychelle Islands).

R. D. Oldham and T. H. Holland, *Records*, Geol. Survey India, vol. 32, pt. 2, 1905, pp. 175–184.

L. L. Fermor, The manganese deposits of India, *Mem.* Geol. Survey India, vol. 37, 1909, pp. 370–380.

G. C. DuBois, Min. pet. Mitt., vol. 22, 1903, p. 1. (Surinam).

A. Lacroix, Nouv. Arch. Mus. Hist. Nat. (Paris), 5th ser., vol. 15, 1913

In apparent contradiction to this many of the worked bauxite deposits are found in temperate regions such as Georgia, Arkansas, France, Hungary, etc., but this is explained by the fact that these are not being formed at the present time but are of Tertiary age when a climate like that of Cuba prevailed in large parts of the temperate zone.

Origin.—The desilication of clay in low latitudes has been discussed extensively. The action of nitric acid, supposedly derived from rain during tropical thunderstorms, has been suggested as the cause. T. H. Holland[1] has mentioned the possibility of bacterial action.

Clay is decomposed by sulphuric acid and by sodium hydroxide or sodium carbonate and at some places aluminum hydroxide may have originated in this way. W. Maxwell[2] has demonstrated this origin for some of the soils of Hawaiian volcanoes and it applies also to a deposit of alum and bauxite on the upper Gila River[3] in New Mexico. Nevertheless it is clear that sulphuric acid does not always produce this effect, for diaspore and hydrargillite occur rarely (Rosita Hills, Colorado; Goldfield, Nevada) in the oxidized portions of mineral deposits where the sericitic rocks are acted upon by sulphuric acid solutions. Bauxite also has rarely been observed. In the oxidized zone the sulphuric acid transforms sericite into kaolin, which is frequently accompanied by more or less alunite ($K_2O.3Al_2O_3.4SO_3+6H_2O$).

These suggestions do not suffice to explain the formation of the lateritic aluminum hydroxides. It is now generally conceded that this is caused simply by the long continued action of ordinary groundwaters under special conditions of moisture and heat. W. J. Mead[4] has shown that there is a complete gradation in case of the Arkansas deposits from the original syenite

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p. 255, reviewed by L. L. Fermor, *Geol. Mag.*, 1915, pp. 28, 77, 123. (French Guinea).

J. M. VanBemmelen, *Zeitschr.* Anorg. Chemie, vol. 66, 1910, p. 322 (General review).

J. Morrow Campbell, Laterite, its origin, structure, etc., *Mining Magazine* (London) Aug.–Nov., 1917. (Tropical Africa.)

[1] T. H. Holland, *Geol. Mag.*, 1903, p. 59.

[2] W. Maxwell, Lavas and soils of the Hawaiian Islands, 1898.

[3] C. W. Hayes, *Bull.* 315. U. S. Geol. Survey, 1906, pp. 215–223.

[4] *Econ. Geol.*, vol. 10, 1915, pp. 28–54.

See also Leith and Mead, Metamorphic geology, New York, 1915, pp. 25–38.

(Fig. 115) that the pisolitic structure develops in place and that residual syenite boulders are surrounded by bauxitic material. The texture of the syenite is sometimes visible in the pisolitic bauxite. There is some evidence of downward leaching of the bauxite, for the top layer is usually more siliceous than the lower parts of the deposit. J. Morrow Campbell believes that bauxite only forms in the zone of percolation close to the fluctuating water level and that it never occurs far below the water.

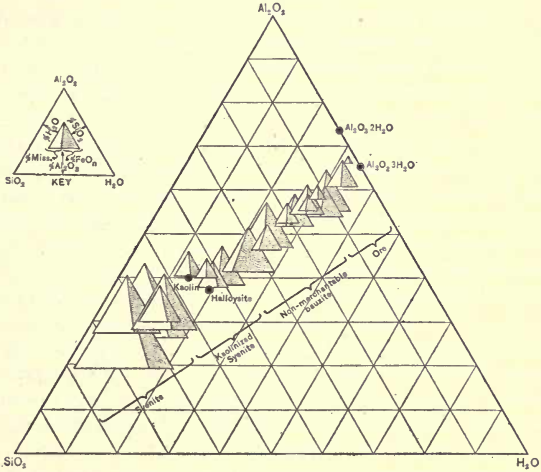

FIG. 115.—Triangular diagram showing the gradation from syenite to bauxite in terms of the principal chemical constituents. Each triangle represents an analyzed sample. *After W. J. Mead.*]

Some bauxites occurring as veinlike masses in limestone of France[1] and Hungary[2] have been explained as the result of the action of ascending waters carrying aluminum sulphate on limestone. Such a mode of origin is admittedly possible.

The sedimentary bauxites of which numerous examples may be found in Georgia and Arkansas in the Cretaceous and Tertiary

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[1] F. Laur, *Trans.*, Am. Inst. Min. Eng., vol. 24, 1894, p. 234.

[2] O. Pauls, *Zeitschr. prakt. Geol.*, vol. 21, 1913, pp. 521–572.

beds are probably deposits swept out into the sea by the normal processes of erosion from bauxite rich laterites on the shore.

Occurrences.—The bauxite deposits of commercial importance are of several different types. In the United States they are confined to Arkansas and the southern Appalachian States.

In Arkansas[1] the mineral occurs in Pulaski and Saline counties as superficial beds over areas of various sizes up to 20 acres. The deposits are only exceptionally more than 10 feet in depth. They rest on nepheline syenite or on a kaolinized form of that rock; the lower part retains traces of granitic structure, while the upper part is distinctly pisolitic. Tertiary sands and clays in places cover the nepheline syenite and the bauxite.

Other deposits of importance, described by Hayes[2] and also by Watson,[3] are found at a number of places in Georgia and Alabama. The principal occurrences are scattered between Jacksonville, Alabama and Cartersville, Georgia, along a belt about 60 miles in length, one of the typical localities being at Rock Run. The bauxite occurs as pockets and irregular masses or curved strata of various colors, with clay and limonite, in the heavy mantle of residual clay overlying the Knox (Cambrian) dolomite, but sharply separated from it. The ore is in part pisolitic and is mined in open cuts, at some places to a depth of 50 feet or more. The bottom of the clay masses is rarely exposed; before it is reached the pockets of bauxite generally terminate in tapering points. Occasionally associated minerals are gibbsite ($Al_2O_3.3H_2O$) and halloysite, which is similar to kaolin in composition but has more water.

A suggestive fact is the occurrence of the deposits at or about the 900-foot contour, which coincides with the elevation of a probable Eocene peneplain. The ores were thus accumulated under topographic and climatic conditions different from those which prevail to-day.

Deposits differing considerably from those already described have recently been found in Randolph and Wilkinson County,

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[1] C. W. Hayes, *Twenty-first Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1901, pp. 435–472.

J. C. Branner, *Jour. Geol.*, vol. 5, 1907, pp. 263–289.

[2] C. W. Hayes, The geological relations of the southern Appalachian bauxite deposits, *Trans.*, Am. Inst. Min. Eng., vol. 24, 1895, pp. 243–254.

W. J. Mead, *op. cit.*

[3] T. L. Watson, *op. cit.*

Georgia.[1] They occur near the contact of the flat-lying sands and clays of the Tuscaloosa (Lower Cretaceous) and Claiborne (Tertiary) formations. The ore occurs either as beds resting directly upon Cretaceous clay or disseminated as nodules through it. A perfect series of transition to clay exists, as shown by analyses. Bauxite beds 10 feet in thickness have been observed; the mineral is clayey, dense, or pisolitic.

Uses and Production.—The annual production of bauxite in the United States has been increasing rapidly and in 1917 was 569,000 long tons, most of which was mined in Arkansas. The mines in France, also yield an increasing amount, about 300,000 long tons in 1913. New deposits are being opened in the British and Dutch Guianas. The bauxite ores contain 35 to 57 per cent. Al_2O_3, a greatly varying percentage of Fe_2O_3, and up to 30 per cent. SiO_2. Ores with more than 4 per cent. Fe_2O_3 are not utilized at present. They are mined in open cuts, often necessitating the removal of heavy overburden, washed to remove the clay, and dried. For purposes of aluminum smelting the ores must be of high grade and low in silica. About 50,000 short tons of aluminum are now produced annually in the United States; exact data are not obtainable. The uses of the metal and its alloys are steadily increasing.

Large works for the electric smelting of aluminum are located at Niagara Falls and in Tennessee. Artificial corundum (alundum) is made from the ore by the electric furnace. Bricks of bauxite for basic non-corrosive lining of furnaces are widely used.[2] Aluminum salts, especially alum, are also manufactured from bauxite. An addition of bauxite promotes the rapid setting of cements.

ANALYSES OF BAUXITE

	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	H_2O	Analyst
Baux, France (pisolitic).	4.8	3.2	55.4	24.8	10.8	Deville.
Jacksonville, Ala.......	21.08	2.52	48.92	2.14	23.41	Hillebrand.
Floyd County, Ga......	0.80	3.52	52.21	13.50	27.72	Nichols.
Pulaski County, Ark....	2.00	3.50	62.05	1.66	30.31	
Wilkinson County, Ga..	9.38	2.76	57.58	0.96	29.12	E. Everhart.

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[1] O. Veatch, *Bull.* 18, Georgia Geol. Survey, 1909, pp. 430–447.

[2] *Mineral Resources*, U. S. Geol. Survey, 1913.

Great variations are often shown in one locality. For further analyses see G. P. Merrill, Non-metallic minerals, 1910, p. 91. The average of a long series of analyses of commercial ore from Georgia tabulated by T. L. Watson[1] gives: SiO_2, 4.274; TiO_2, 3.791; Al_2O_3, 58.622; Fe_2O_3, 1.507; and H_2O, 31.435; total 99.629. This corresponds to $Al_2O_3.3H_2O$.

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[1] *Bull.* 11, Georgia Geol. Survey, 1904, pp. 45–46.

CHAPTER XIX

THE HEMATITE DEPOSITS OF THE LAKE SUPERIOR REGION

General Character, Distribution.—The iron-ores mined in the Lake Superior region in Minnesota, Michigan and Wisconsin amount to from 80 to 90 per cent. of the total domestic output and in 1917 yielded 64,000,000 long tons. The ore is mainly hematite with small admixtures of limonite and magnetite. It occurs as masses, lenses, or flat deposits in pre-Cambrian sedimentary rocks of Algonkian and Archean age. The deposits are concentrated by the oxidizing and silica-dissolving effect of waters of meteoric origin, in original sediments called "iron formation" which were rich in carbonate and silicate of iron. They are products of pre-Cambrian weathering which, probably under arid conditions, reached depths not approached elsewhere. Only to a small degree and near the surface does this ore forming activity of the waters persist at the present time.

We owe most of our information concerning these deposits to the work of C. R. Van Hise, C. K. Leith, and many others recorded in a magnificent series of monographs of the United States Geological Survey. These and other papers are cited below.[1]

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[1] R. D. Irving and C. R. Van Hise (Penokee district) *Mon.* 19, U. S. Geol. Survey, 1892.

C. R. Van Hise and W. S. Bayley (Marquette district) *Mon.* 28, 1897.

J. M. Clements and H. L. Smyth (Crystal Falls district) *Mon.* 36, 1899.

C. K. Leith (Mesabi district) *Mon.* 43, 1903.

J. M. Clements (Vermilion district) *Mon.* 45, 1903.

W. S. Bayley (Menominee district) *Mon.* 46, 1904.

C. R. Van Hise, Iron ore deposits of the Lake Superior region, *Twenty-first Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1901, pp. 305–434.

C. R. Van Hise and C. K. Leith, The geology of the Lake Superior region, *Mon.* 52, 1911.

S. Weidman, The Baraboo iron-bearing district, Wisconsin, *Bull.* 13, Wisconsin Geol. and Nat. Hist. Survey, 1904.

C. K. Leith, A summary of Lake Superior geology, *Trans.*, Am. Inst. Min. Eng., vol. 36, 1906, pp. 101–153.

C. K. Leith, The geology of the Cuyuna iron range, Minnesota, *Econ.*

There are seven principal districts in the United States and three or four in Canada, locally called ranges, as follows (Fig. 116):

1. The Mesabi, Vermilion, and Cuyuna ranges of northern Minnesota.

2. The Penokee-Gogebic, Marquette, Iron River and Menominee ranges, mainly in northern Michigan.

3. The Baraboo range of southern Wisconsin.

4. The Michipicoten, Gunflint Lake, and other minor districts in Canada.

Geology.—The following short summary is in part taken from the résumés by Leith and Harder cited above.[1]

The principal rocks of the Lake Superior iron ore region are divided as follows:

Cambrian: Potsdam sandstone.
Algonkian: Keweenawan series (sediments, basic flows, gabbro,
Huronian series:
  Upper Huronian (quartzite, "iron formation," and slate).
  Middle Huronian (quartzite, "iron formation," and slate).
  Lower Huronian (quartzite, conglomerate, dolomite, slate, "iron formation," and intrusives).
Archean: Laurentian series (granite, gneiss, and porphyry).
Keewatin series (greenstone, amphibolite, and "iron formation").

Geol., vol. 2, 1907, p. 145; The iron ores of Canada, *Econ. Geol.*, vol. 3, 1908, pp. 276–291.

E. C. Harder and A. W. Johnston, Notes on the geology of the Cuyuna district, *Bull.* 660, U. S. Geol. Survey, 1917, pp. 1–26.

J. F. Wolff (Mesabi range), *Eng. and Min. Jour.*, July 17–Aug. 7, 1914; *Trans.* Am. Inst. Min. Eng., vol. 56, 1917, pp. 142–169.

E. C. Harder, *Mineral Resources*, U. S. Geol. Survey, 1908, Summary.

E. F. Burchard, The production of iron ore, etc., *Mineral Resources*, Annual publication.

Carl Zapffe, The Cuyuna iron ore district, Brainerd Tribune, *Suppl.*, July 1, 1911.

[1] Classifications differing somewhat from that here given have been proposed in Canada by Coleman, Miller and Knight, but for present purposes the nomenclature current in the United States is adhered to. See W. G. Miller and C. W. Knight, *Bull.* Geol. Soc. Am., vol. 26, 1915, p. 87; *Rept.* Ontario Bur. Mines, vol. 22, pt. 2, 1914.

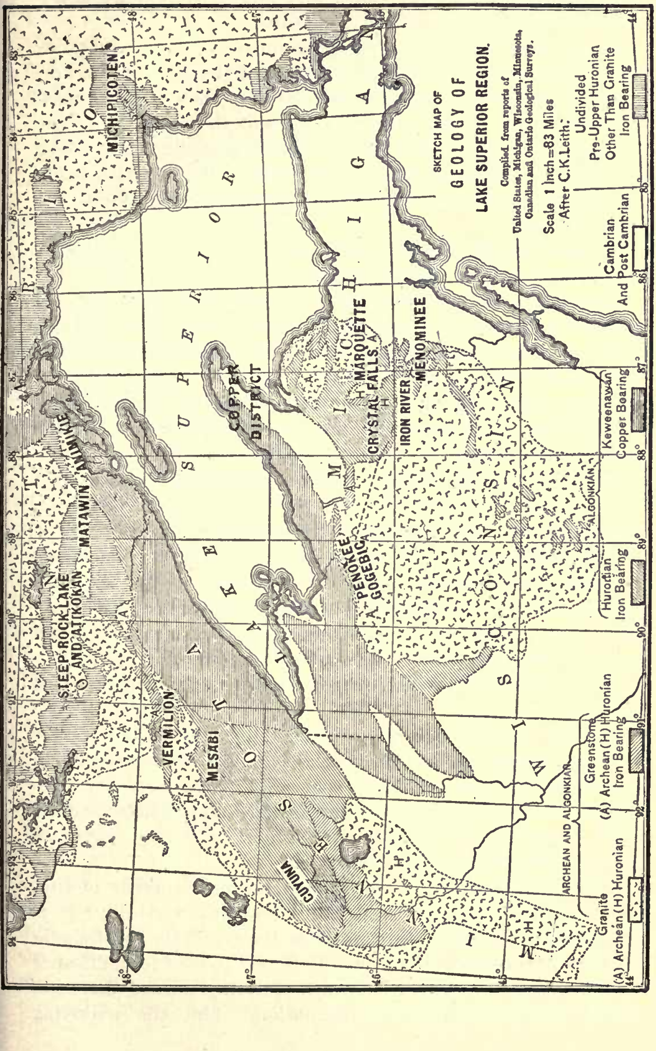

SKETCH MAP OF
GEOLOGY OF
LAKE SUPERIOR REGION.
Compiled from reports of
United States, Michigan, Wisconsin, Minnesota,
Canadian and Ontario Geological Surveys.
Scale 1 Inch=83 Miles
After C.K.Leith.
Undivided
Pre-Upper Huronian
Other Than Granite
Iron Bearing
Cambrian
And Post Cambrian
Keweenawan
Copper Bearing
Huronian
Iron Bearing
Greenstone
(A) Archean (H) Huronian
Iron Bearing
Granite
(A) Archean (H) Huronian
ARCHEAN AND ALGONKIAN
ALGONKIAN
LAKE SUPERIOR
ONTARIO
MICHIGAN
WISCONSIN
MINNESOTA
MICHIPICOTEN
STEEP ROCK LAKE
AND ATIKOKAN
MATAWIN
ANIMIKIE
VERMILION
MESABI
CUYUNA
COPPER DISTRICT
PENOKEE
GOGEBIC
MARQUETTE
CRYSTAL FALLS
IRON RIVER
MENOMINEE
48°
47°
46°
45°
44°
83
84
85
86
87
88
89
90
91
92
93
94

FIG. 116.

Of these rocks only the Upper and Middle Huronian and the Keewatin contain deposits of hematite.

The Archean or basement complex consists of gneiss and granite with an extensive series of greenstones (basalt, gabbro, amphibolite), which are largely surface lavas. These lavas are now regarded as the oldest formation exposed; the character of the basement upon which they were outpoured is unknown. Above the Keewatin lavas lie sedimentary rocks of the iron formation. The gneisses and granites are in part certainly intrusive into the Keewatin series.

In the Vermilion and Michipicoten districts the productive formation is in the Keewatin series.

Unconformably overlying the Archean and similarly covered by the Cambrian is the Algonkian, which in its complete development consists of four parts separated by uncomformities. The lower three divisions are collectively referred to as the Huronian and the uppermost as the Keweenawan. The principal iron-bearing formations are concentrated in the Huronian, but the development differs materially in the several districts.

In the Marquette district all three divisions of the Huronian are present. The lower Marquette series consists of quartzite, dolomite, and slate 3,000 feet in maximum thickness. The middle Marquette series, 3,000 feet in maximum thickness, includes quartzite, slate, and the important Negaunee iron-bearing formation. The upper Marquette series includes quartzite, schist, slates, and fragmental basic volcanic rocks, each member accompanied by iron-bearing formations.

In the Crystal Falls and Menominee districts similar divisions appear.

In the Penokee-Gogebic district the Upper and Lower Huronian series are present, but the middle series appears to be lacking. The lower division consists of quartzite, cherty limestone, and dolomite; the upper part includes quartz slates and the thick Ironwood "iron-bearing formation" and may aggregate about 13,000 feet in thickness.

In the Mesabi district the Lower Huronian consists of conglomerates, graywackes, and slates standing vertically; it is intruded by the granite of the Giants Range, on the south slopes of which the iron deposits extend from east to west for a distance of 100 miles. The Upper Huronian comprises a basal quartzite, the Biwabik "iron-bearing formation," and the overlying

Virginia slate. The total thickness is probably over 2,000 feet. The series dips gently at angles of 5° to 20° and is also gently cross folded. Intrusive into these rocks at the east end of the district are Keweenawan granite and basic igneous rocks. Near these intrusives the sedimentary rocks are highly metamorphosed.

Above the Huronian rests the less highly metamorphosed Keweenawan series of sandstones, conglomerates, and igneous basic flows; the thickness is estimated to be over 35,000 feet. It contains no iron deposits.

The west end of Lake Superior consists of an eastward-pitching synclinorium of Keweenawan rocks. The next underlying series, the Upper Huronian, takes less part in this synclinal structure and borders the other edge of the Keweenawan areas.

We have thus in the Lake Superior country six series, consisting from top to bottom of the Keweenawan, Upper, Middle, and Lower Huronian, Laurentian, and Keewatin, all but the last two separated by unconformities. Above them and separated by a marked unconformity rests the Cambrian Potsdam sandstone.

The "Iron Formations."—The iron ores of the Lake Superior region are believed to be derived by concentration by means of meteoric waters from lean "iron formations" containing about 25 per cent. iron. The ores are products of enrichment of chemically deposited sediments, such as siderite and hydrated iron silicates, for the most part interbedded with normal clastic sediments, such as slate and quartzite.

The iron formations range from a few feet up to 1,000 feet in thickness and are sedimentary beds consisting, according to Leith, "mainly of chert, or quartz, and ferric oxide segregated in bands or sheets, or irregularly mingled. Where in bands with the quartz layers colored red and the rock highly crystalline it is called jasper. Where less crystalline and either in bands or irregularly intermingled the rock is known as ferruginous chert. The silica in these rocks varies from 32 to 80 per cent., the ferric oxide from 31 to 66 per cent. Other phases of the iron formation, subordinate in quantity, are (1) ordinary clay slates, showing every possible gradation through ferruginous slates into ferruginous cherts; (2) paint rocks, oxidized equivalents of the slates; (3) cherty iron carbonate (siderite) and hydrous ferrous silicate (greenalite); (4) the iron ores themselves. Almost the entire bulk of the iron formations now consists of iron oxide and silica, with carbonates and alumina present in subordinate quantity."

Spurr and Leith found that certain rocks of the Mesabi district contained, in a matrix of chert and iron carbonate, abundant round granules of a green chloritic substance which Leith called greenalite; its composition is approximately 30 to 38 per cent. SiO_2, 8 to 34 per cent. Fe_2O_3, 25 to 47 per cent. FeO, and 7 to 9 per cent. H_2O (p. 263). The absence of potassium shows that the mineral is not glauconite. The greenalite rocks contain 50 to 80 per cent. of this mineral, which is soluble in acids. The same mineral occurs in some of the siderite rocks of the more easterly districts.

Regional metamorphism and, to a greater degree, contact metamorphism, caused by Keweenawan intrusions of granites and gabbros, have, in places, converted the siderite and the greenalite rocks to magnetite-amphibole schists and the soft hematite to specularite; this is especially well observed in the Marquette and the Mesabi ranges.

The Iron Ores.—The hematite ores are derived from the ferruginous cherts by a process of concentration, and both laterally and in depth gradually change into such rocks. The ores are admixed with enough magnetite to affect the magnetic needle and render possible magnetic surveys of the fields. The hard blue specular ores of the Marquette range contain more magnetite than the others and are accompanied by contact-metamorphic jaspers and magnetite-amphibole (grünerite) rocks. In other ranges, such as the Mesabi, Penokee, and Baraboo, the ore is soft, bluish, red, or brown in color, and partly hydrated. A micaceous or foliated development of the iron ore is not common. The average analysis of Lake Superior ores in 1909 is as follows:

	Per cent.
Moisture (loss at 100° C.)	11.28
Analysis of dried ore:	
Iron	58.45
Phosphorus	0.091
Silica	7.67
Alumina	2.23
Manganese	0.71
Lime	0.54
Magnesia	0.55
Sulphur	0.06
Loss by ignition	4.12

This corresponds to a composition as follows:

Hematite (more or less hydrated)	86.45
Quartz	4.89
Kaolin	5.25
Chlorite	1.01
Dolomite	0.81
Apatite	0.48
Miscellaneous	1.11
	100.00

The tenor in iron of the shipped ore has slowly diminished during recent years; in 1905 it was 59.6 per cent. Fe.

The phosphorus ranges between 0.008 and 1.29, the bulk of the ore being of Bessemer grade—that is, containing less than 0.05 per cent. phosphorus. Small parts of the ore shipped, particularly from the Mesabi and Cuyuna ranges contain as much as 7 per cent. manganese. The sulphur varied from 0.003 to 1.87 per cent., but it averages low. Accessory, more or less rare minerals in the ore, aside from quartz or chert, are apatite, wavellite, adularia, calcite, dolomite, siderite, pyrite, marcasite, chalcopyrite, tourmaline, ottrelite, chlorite, garnet, mica, rhodochrosite, barite, gypsum, analcite, goethite, and turgite.

The ore reserves of the Mesabi range are estimated to be 1,385,000,000 tons; those of the whole region 1,475,000,000 tons.

The total yield of the Lake Superior ores from 1854 to 1916 has been about 770,000,000 long tons, much the greater proportion having been extracted in the last three decades.

Carbonate ores are now mined in the Michipicoten district, Canada.

Form of Ore Bodies.—The ore forms irregular, often very large, but as a rule distinctly bedded or banded masses in the "iron formations;" in places it is entirely embedded in them. The shape is commonly determined by impervious basements like clayey dikes, decomposed amphibolitic rocks, or folded sedimentary beds like slate, which have tended to guide the circulation of surface water into certain channels; the ores usually occur in pitching troughs caused by any or all of these factors.

In some ranges like the Gogebic, Marquette, and Iron River the strata are strongly folded and may dip at high angles; some of the ore-bodies have been followed to a depth of 1,500 or 2,000 feet. Good ore is mined at present in the Newport mine in the Gogebic district at 2,000 feet. In the Mesabi range the rocks lie horizontal; the alteration and concentration have extended over a wide area and few of the mines are deeper than 200 feet. The shallow deposits of this range are mined on an enormous scale by steam-shovels. The annual production, which reached 41,127,323 long tons in 1917, is far greater than that of other districts.

Marquette Range.—The mines of the Marquette range are near Negaunee and Republic, south and southwest of Marquette. The principal "iron formation," the Negaunee, is in the Middle Huronian, and the sedimentary rocks are intruded and metamorphosed by basic igneous rocks. Extensive folding has taken place and the strata are compressed into a great synclinal basin.

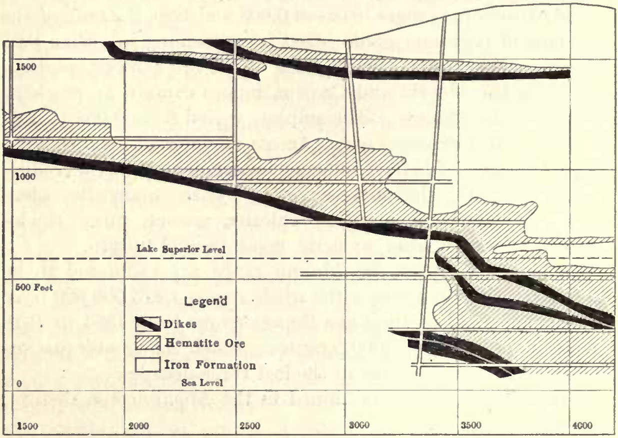


Fig. 117.—Longitudinal section of the Montreal Mine, Gogebic Range, Michigan, showing dependence of bodies of oxidized iron ore on dikes.[1]

The ores lie at the base of the Negaunee formation; where the underlying slates have been folded so as to form pitching synclinal basins, or where dikes have guided the concentrating waters. In part they occur also at the contact of the iron formation with basic intrusions—for instance, in pitching troughs between igneous masses and dikes branching from them. The surfaces of the igneous rocks are much altered, leached, and changed to clayey masses, called "soapstone" and "paint rock."

Menominee Range.—The iron-bearing district extends from western Michigan into Wisconsin, the principal mines being located at Iron Mountain, Norway, and Crystal Falls. The iron formation is chiefly in the Upper Huronian and is called the Vulcan formation; it is overlain by Upper Huronian slate and underlain by a Lower Huronian dolomite. Intricate folding characterizes the structure of the range, the ores of the different areas occurring in separate local basins. The deposits are large and consist of soft red hematite, considerably hydrated in places, and are generally found in pitching synclinal basins bottomed and capped by slate layers.

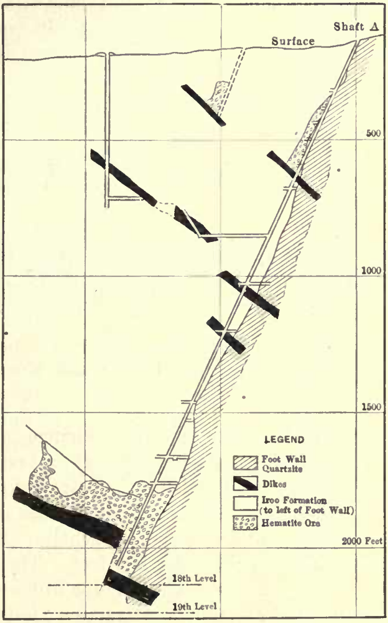


FIG. 118.—Vertical cross-section of the Newport Mine, Gogebic Range, Michigan, showing position of ore-bodies above dikes. *Data from H. L. Smyth.*

Penokee-Gogebic Range.—This range is in northern Michigan and Wisconsin, the principal mines being at Hurley, Ironwood, and Bessemer. The ore appears in the Ironwood formation (Upper Huronian), which is overlain by slate and underlain by quartzite and black slate. The dip is steep and the sediments are in part metamorphosed by Keweenawan gabbro; for the most part the Ironwood formation is ferruginous chert. The ores are concentrated in large irregular bodies in the angles between the footwall quartzite or black slate and the igneous dikes (Figs. 117 and 118), these rocks making an impervious trough, toward which the meteoric waters converged. Most of the deposits reach depths of 1,000 feet, and some attain 2,200 feet. Both soft, partly hydrated ore and hard slaty ore occur.

Cuyuna Range.—The Cuyuna district is situated near Brainerd about 70 miles southwest of the Mesabi mines. It extends for 65 miles along the strike of the rocks in a northeast direction. The iron ore here is a partly hydrated hematite, in places accompanied by an unusual amount of manganese oxide (up to 30 per cent. Mn). It is contained in the usual iron formation of ferruginous jasper which in depth appears to change to cherty iron carbonate. The enclosing rocks are slates of various kinds compressed into steep folds, the details of which are difficult to trace owing to the covering glacial drift. The ore bodies are elongated following the strike and while some cease at shallow depths others have so far been followed down for 300 feet. The phosphorus ranges from 0.1 to 0.5 per cent. The Cuyuna iron ores were discovered by means of the magnetic attraction along the range, due to a small quantity of admixed magnetite.

FIG. 119.—Generalized cross-section showing relation of iron-bearing formation to associated rocks in the Mesabi Range, Minn. *After J. F. Wolff.*

Mesabi Range.—In northern Minnesota the Mesabi range extends from east to west for a distance of 75 to 100 miles on the south slope of a prominent ridge called the Giants Range. The principal mines are situated near the towns of Biwabik, Eveleth, Virginia, and Hibbing. The Huronian rocks here lie at gentler inclinations than elsewhere, dipping 8° to 10° S.E. so that the iron formation outcrops in a general northeast-southwest belt (Fig. 119).

The Biwabik iron formation of the Upper Huronian contains the deposits. It is underlain by the Pokegama quartzite and covered by the thick Virginia slate, a chloritic and aluminous sedimentary rock. Except at the eastern end of the range, where contact-metamorphic amphibole-magnetite rocks have developed, the iron formation is composed mainly of ferruginous chert. The iron ores cover large irregular areas along the outcrop of the Biwabik formation, but descend to relatively slight depths, few of the mines being more than 200 feet deep (Fig. 120). The deposits are most abundant at the synclines of the transverse folds of the formation. They are bedded and along the edges change rather abruptly to the ferruginous chert, from which they are derived by leaching of the silica. This relationship is clearly indicated by the slumping of the strata near the edges of the ore masses, as shown in Fig. 121. The iron formation is locally called "taconite."

The rain water falling on the truncated edges of the beds circulates toward the south, its movement being controlled by the slight synclinal basins, by impervious layers of slate, and by fractures.

The secondary concentration of the iron ore has evidently taken place under surface con-

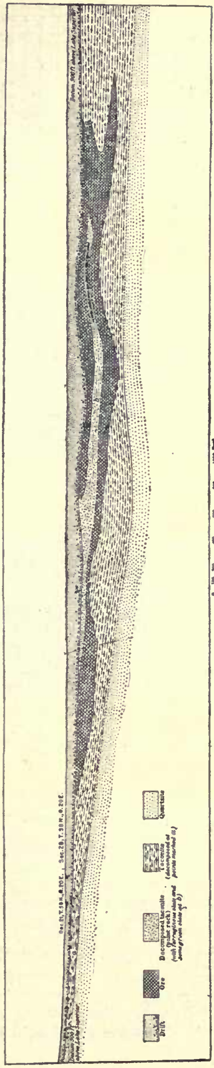
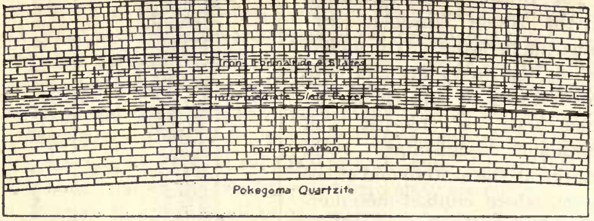
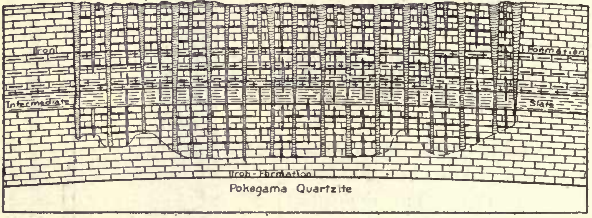


FIG. 120.—North-south cross-section through iron-bearing Biwabik formation, Mesabi district, Minnesota. *After O. B. Warren and C. K. Leith, U. S. Geol. Survey.*

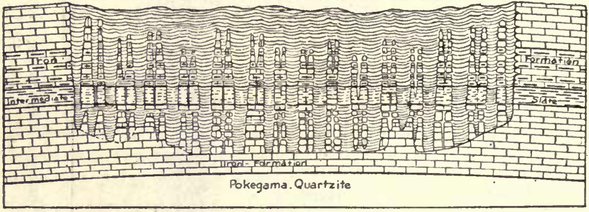
A-Tension Cracks in Iron-Formation on Axis of an Anticline.



B-Ore forming by alteration of Taconite along Fissures & Bedding-Planes.



C-Alteration of Taconite to Ore well advanced.



D-Present Condition of Average Trough Orebody.

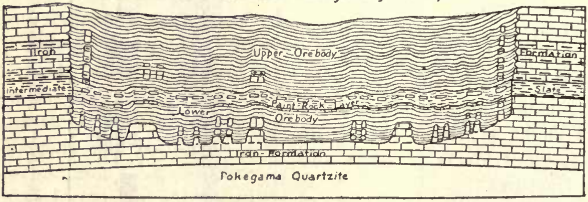


FIG. 121.—Cross-section showing mode of development and slumping of ore-body at Mesabi Range. *After J. F. Wolff.*

ditions since the remote time of the post-Keweenawan folding, when the deposits first became exposed; it has also taken place below as well as above the present water-level, which is about 75 feet underneath the surface.

Analyses show that the present surface water, containing about 20 parts per million of SiO_2, is slowly leaching silica, but removes little if any iron. The deposits do not appear to continue underneath the edge of the capping Virginia slate, probably because of the ponding of the water below that impervious formation. The amphibole-magnetite rocks in the eastern part of the district are more stable and have not suffered much alteration by oxidation.

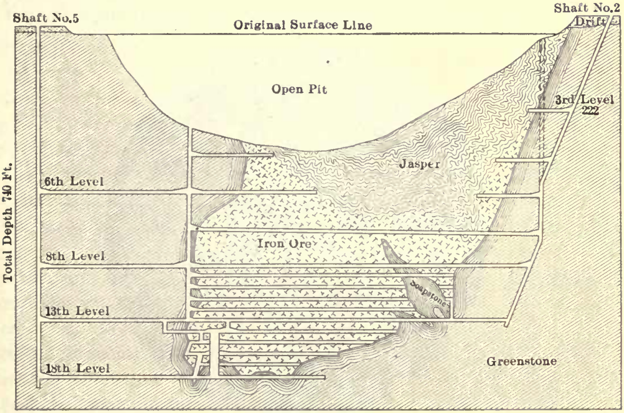


FIG. 122.—Vertical section through the Chandler mine, Vermilion range, Minnesota. *After J. M. Clements, U. S. Geol. Survey.*

During the development of the ore-bodies erosion has continually cut down the iron formation and this truncation has been accompanied by slow downward and lateral migration of the iron. Glacial erosion finally removed much material.

The ore is a soft and porous hematite, brown, red, or blue in color, averaging 55 to 58 per cent. iron. It contains little magnetite, but some turgite and goethite. The mineral composition of the ore in 1909 was approximately in per cent.: hematite, 61.81;

limonite, 25.95; quartz, 4.10; kaolin, 5.30; manganese dioxide, 1.30; miscellaneous, 1.54.

Sulphur is low and phosphorus varies from 0.03 to 0.07 per cent. There is considerable more phosphorus in the ore than in the ferruginous chert; the greenalite and siderite rocks contain scarcely any phosphorus.

Vermilion Range.—Northeast of Mesabi, near the Canadian boundary, is the Vermilion range, the principal mines being near the towns of Ely and Tower. The country rock is mostly the Keewatin greenstone, but infolded in it in synclinal basins or troughs is the Laurentian iron formation, known as the Soudan. The ores are associated with ferruginous jaspers in these troughs and generally have a footwall of greenstone (Fig. 122). The ore is a dense and hard blue or red hematite which contains a little chalcopyrite, an unusual feature in this region.

Fig. 123.—Ferruginous chert with greenalite granules, in part replaced by ferric oxide (black). Magnified 40 diameters. *After C. K. Leith.*

Origin of Lake Superior Iron Ores.—It has been shown by Van Hise and Leith and their associates that the ferruginous cherts, jaspers, amphibolite-magnetite schists, and iron ores of the iron formations result from the alteration either of the cherty iron carbonate or of the greenalite. The small amounts of iron carbonate or ferrous silicate now found in the formations represent mere remnants left unaltered where protected by other rocks. The steps of the alteration may be observed and, in the end products, the structures and textures of the original rock are often remarkably well retained. It is held that the ores and the ferruginous cherts or jaspers on one hand and the amphibole schists on the other hand represent alterations from the same original type. The source of the ore is not, as a rule, in the present ferruginous cherts, but it was developed from original lean siderite and greenalite rocks. It is held that in the largest deposits ores and jaspers may have developed side by side, at the same time, from such original minerals. Iron carbonate prevailed in the Marquette, Gogebic, Vermilion, and Crystal Falls districts; greenalite in the Mesabi district (Fig. 123).

The concentration has been effected, according to the Lake Superior geologists, by water coming more or less directly from the surface, especially at places where such waters converge owing to the existence of impervious underlying formations, such as slate or "soapstone," that form pitching troughs, or owing to brecciation and fracturing of the iron formations.

The alteration of the iron formations, resulting in the concentration of the iron ores or in the development of ferruginous cherts, jaspers, and amphibolite schists, has taken place in different geologic periods under varying conditions. So far as the alteration has proceeded continuously under the influence of surface waters, without interruption by igneous activity or orogenic movements, soft ores and ferruginous cherts have resulted. So far as these products have been subjected to deep-seated alteration they have become dehydrated into hard red and blue specular ores and brilliant jaspers. So far as the alteration of the original iron formations has taken place within the sphere of influence of great intrusive masses, when waters were heated and oxygen not abundant, or under similar conditions, developed by deep submergence or by orogenic movement, ferrous silicates and magnetite resulted, as shown in the development of the grünerite schists.

The concentration of the ores was far advanced before Cambrian time, as shown by the fragments of ores in Cambrian conglomerates. Most of the deposits were formed between the Keweenawan and the Cambrian deposition. At the close of pre-Cambrian time the ores were largely as we now find them, though some concentration has been going on since. During the Cretaceous period the region of the Mesabi range, at least, was covered by the sea.

Regarding the origin of the cherty iron carbonates, Van Hise has held that they were derived largely from the more ancient basic volcanic rocks of the Lake Superior region. The iron was leached by underground waters and carried to the sea as carbonate, partly also as sulphate solution, and there deposited as limonite, from which through reduction by organic matter ferrous carbonate was formed.

Somewhat different views have lately been expressed by C. K. Leith,[1] who sums up the origin of the ores as follows: The iron was brought to the surface by igneous rocks and either contributed directly to the ocean by hot magmatic waters or later brought there by surface waters from weathered rocks. The iron-bearing minerals were then deposited as a chemical sediment in a conformable succession of sedimentary rocks and still later, under conditions of weathering, were locally enriched to ore by percolating surface waters. "It begins also to appear that the iron, copper, nickel, and silver ores of the Lake Superior and Lake Huron districts are related in a great metallographic province in which the characteristics and distribution of the different ores are initially controlled by igneous rocks. As first deposited the iron formation consisted essentially of iron carbonate or ferrous silicate (greenalite) with some ferric oxide, all minutely interlayered with chert, forming the ferruginous chert. When these were exposed to weathering the ferrous compounds, the siderite and greenalite, oxidized to hematite and limonite, essentially *in situ*, although some of it was simultaneously carried and redeposited. The result was ferruginous chert or jasper, averaging less than 30 per cent. of iron. The concentration of the iron to 50 per cent. and over has been accomplished essentially by the leaching of silica bands from the ferruginous chert and jasper. Infiltration of iron has been on a smaller and more variable scale. The leaching of the silica develops pore space and allows the iron layers to slump, thereby enriching the formation sufficiently to constitute an ore." Only a small part of the volume of the iron formations—less than 2 per cent.—has been altered to ore.

Résumé.—The literature of the Lake Superior iron ores is extensive and many different views have been expressed. J. D. Whitney regarded the ores as of igneous origin, and this view has also been advocated by N. H. Winchell. T. B. Brooks and R. Pumpelly at one time considered them as dehydrated bog iron

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[1] C. K. Leith, Iron ores of Canada, *Econ. Geol.*, vol. 3, 1908, pp. 276–291.

ores, and this view has lately been adopted by S. Weidman in his description of the Baraboo ores of Wisconsin, where the ores appear to grade into dolomites.

The views of Van Hise and Leith and their associates, which appear to be generally accepted, have been given above in some detail and in part verbatim. The development of their theory of the origin of the iron ores has been gradual; at first the iron formations were considered as purely sedimentary and the recrystallization to amphibole-magnetite rocks as evidence of regional metamorphism; later the effects of contact metamorphism were recognized, and finally it is held that the iron of the iron formations was in large part yielded by the extensive eruptions accompanying their deposition. Although the ferruginous cherts are still thought to be formed by the oxidation of the siderite and greenalite rocks, which now form a small part of the formations, there seems to be a tendency to regard the iron ores as mainly formed directly by the solution of the silica in the ferruginous cherts. The history of any one group of these deposits is probably even more complicated than would appear from the descriptions. In the Mesabi range, for instance, amphibole and adularia occur in the ore, but the development of both these minerals is incompatible with descending and oxidizing waters.

The age of the concentration of the iron deserves emphasis. The ores were formed mainly before the Cambrian, as indicated by fragments of ore in the Cambrian conglomerate. Indeed, they were in part developed in inter-Huronian time, even in early Huronian time. This is set forth in the publications cited, but is not generally realized. The ores are not the product of the present circulation and oxidation, but of forces acting in ancient periods when conditions were probably widely different from those of to-day. It is stated that the concentration has also proceeded since pre-Cambrian time, but this assertion seems far from being established, even for the surface deposits of the Mesabi range.

Weidman (*op. cit.*) has pointed out that the present groundwaters are entirely similar in composition in the Paleozoic rocks and in the iron formations and has shown that they do not now transport or dissolve iron or notable quantities of silica. A. C. Lane has shown (p. 440) that the depth reached by the potable surface waters is limited and that in some parts of the iron districts, as well as in the copper districts of the Keweenawan, they are replaced, at depths of 1,000 to 2,000 feet, by scant and apparently stagnant water rich in calcium and sodium chlorides. The present ground-water in a region of high water-level is clearly unable to produce the extensive oxidation shown by the iron ores. Undoubtedly special conditions of circulation existed in pre-Cambrian time which are not paralleled to-day. Oxidizing waters do not penetrate deeply in temperate regions of high water-level, and even where they reach a depth of a few hundred feet the product is a limonite. The hematites appear to result from oxidation only in arid and tropical countries.

The only times at which large bodies of rock could be oxidized to hematite by descending waters would seem to be during epochs of great aridity, when the water-level was exceptionally low; possibly just such conditions prevailed in pre-Potsdam time. Similar extraordinary deep oxidation of pre-Cambrian age has been described by L. L. Fermor from the manganese deposits of India.[1] It may be pointed out that deposits of hematite were formed during this period in the Hartville district in eastern Wyoming,[2] and finally in certain recently described areas in western Arizona.[3]

At Hartville lenses of hematite occur in schist along a limestone footwall and have been followed to a depth of 900 feet. Ball shows that the deposit antedates the Guernsey formation, the lowest Paleozoic terrane present, and believes that the iron was leached by descending solutions from the upper part of the schist and deposited in its lower part by replacement.

Another question of possible importance relates to the percentage of phosphorus in the Lake Superior ores. It is remarkably low for sedimentary deposits in the origin of which organic life played a part. It is still more remarkable that the primary greenalite rocks at Mesabi are almost free from phosphorus.

The Baraboo deposits of Wisconsin are peculiar in that igneous rocks are there entirely absent, and in that the hematite grades into the overlying dolomite; the argument advanced by S. Weidman in favor of a primary deposition of the ore as limonite or hematite is not without strength.

In spite of the great amount of work done the problem of the origin of the Lake Superior hematites still possesses some puzzling features.

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[1] *Mem.*, Geol. Survey India, vol. 37, 1909.

[2] S. H. Ball, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 190–205.

[3] H. Bancroft, *Bull.* 451, U. S. Geol. Survey, 1911.

CHAPTER XX

DEPOSITS FORMED BY CONCENTRATION OF SUBSTANCES CONTAINED IN THE SURROUNDING ROCKS, BY MEANS OF CIRCULATING WATERS

GENERAL STATEMENT

The water which sinks through the soil and effects the weathering of rocks becomes charged with small amounts of carbonates of calcium, sodium, magnesium, potassium, iron, and other metals, and also with soluble silica. By far the larger part of it, after a short journey through the belt of oxidation, returns to the surface as springs and seepage and is carried off in the watercourses to the sea. A smaller part of this water sinks into the ground and either joins the active circulation, descending in smaller fractures and openings to ascend on the larger fissures and other waterways, or becomes a part of the stagnant or almost stagnant and gradually diminishing ground-water of deeper levels. In places the active circulation may descend to depths of 10,000 feet. In comparison with the depth of the ground-water, the depth of oxidation or rock decay is on the whole insignificant, and that part of the dissolved substance which is carried down is also insignificant in comparison with the vast amount of underlying rocks, so that we cannot expect that the material added from the zone of weathering will produce any far-reaching changes in the composition of these rocks.

Nevertheless dissolved salts are carried down from the weathered belt and may cause deposits in open cavities or may form more or less complex replacements. In the openings silica may be deposited as chalcedony, chert, or quartz; calcium carbonate may fill fissures and replace silicates, or ferrous carbonate may be substituted for limestone. Chlorite, kaolin, and sericite may develop in igneous rocks. All these changes are, however, accompanied by renewed solution, and it is a debatable question whether the solution does not more than balance deposition. On the other hand, the water returning to the surface after a journey of varying length, more or less heavily loaded with soluble salts deposits these by reason of decrease of temperature or by reaction with other surface waters of different composition. Finally, hydration absorbs much water, both from the active circulation and from the more stagnant ground-water, and deposits of valuable minerals may result from this simple process.

In a rough way the deposits resulting from the work of underground waters of meteoric origin may be divided into (1) those formed from abundant material contained in the surrounding rocks, for instance, magnesite, serpentine, sulphur (by reduction of gypsum), and certain kinds of hematite; and (2) those formed by the deposition of rarer substances dissolved by the water from the surrounding rocks or from rocks that lie deeper. In this second division it is possible to indicate with great confidence the derivation of some substances—*e.g.*, barite from certain limestones, and copper from certain basic igneous rocks; but the exact derivation of some other substances may be doubtful.

Waters of atmospheric origin doubtless have the power to dissolve many of the rarer metals contained in rocks, to carry them for considerable distances, and to concentrate them in places suitable for deposition; but unless it is aided by higher temperatures at considerable depths below the surface this power is probably not strong enough to produce important deposits of these rarer metals.

BARITE[1]

Modes of Occurrence and Origin.—Barite, the sulphate of barium, also known as barytes or heavy spar, contains when pure 65.7 per cent. BaO and 34.3 per cent. SO_3. It is usually white and coarsely crystalline with curved cleavage faces but appears also, especially in residual deposits, with granular, earthy or even fibrous texture. Many barites contain from a fraction to several per cent. of strontium sulphate; the material mined is often quite pure except for small amounts of silica, calcite, gypsum, kaolin and iron hydroxide.

Witherite, the barium carbonate, is a much rarer mineral and is found in barite veins, associated with galena. It occurs rather abundantly in such veins in Cumberland and Northumberland

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[1] E. F. Burchard and W. C. Phalen, *Mineral Resources*, U. S. Geol. Survey, Annual publication.

J. M. Hill, Barytes and Strontium. An excellent review in same publication for 1915, pt. 2, pp. 161–187.

Mineral Industry, Annual publication.

in England, particularly at the New Brancepeth colliery where it is secondary after barite.

Barite is not a mineral of igneous origin, nor does it occur in contact metamorphic deposits. It is common, however, as lenses and veins in almost all kinds of rocks. Most abundantly, however, it occurs in sedimentary rocks and is probably in almost all cases leached from the country rock by circulating waters.

All igneous rocks contain at least a trace of barium oxide but rarely more than 0.1 per cent. Leucite and analcite rocks from the Wyoming-Montana province are unusually rich in this metal, some analyses showing from 0.5 to 1.2 per cent. of BaO probably present in the feldspathoids as a silicate. Granites, rhyolites, andesites, and basalts are poor in barium. In more concentrated form barium is often present in sediments. It has been determined in some limestones; analyses quoted by T. L. Watson from the Ordovician of Virginia show from 0.62 to 1.62 per cent. $BaSO_4$.[1] Limestones of the same age from Missouri contain, according to Steel, only 0.001 to 0.005 per cent. BaO. Some sandstones, like the Cambrian Weisner quartzite of Georgia and Alabama contain barite; it has also been found in shales with sedimentary manganese ores (p. 274). Sea water contains traces of barium and strontium and many natural waters particularly salt brines, hold quite a little barium in solution as chloride or carbonate. Many cases of deposition of barite by natural waters are mentioned on pp. 103–108.

Barium sulphate is soluble in water to the extent of 2.9 milligrams per liter; it is somewhat more soluble, probably with decomposition, in waters containing alkaline carbonates and chlorides.

Barite is a common gangue mineral in many ore deposits but is here scarcely ever of economic importance. In most cases its origin is to be sought in the rocks traversed by the ascending solutions. The barite deposits worked generally contain few other minerals and are found in sedimentary rocks of all ages as veins and lenses whose width in places may be from 10 to 50 feet. Sometimes a little pyrite, galena and sphalerite is asso-

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[1] C. W. Dickson, The concentration of barium in limestone, *School of Mines Quart.*, vol. 23, 1902, pp. 366–370. This author fails to find it in many limestones. F. W. Clarke does not mention its presence in single and composite analyses of limestone, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 558.

ciated with the barite. The source of the mineral is undoubtedly in the surrounding sediments from which it has been dissolved by meteoric waters.

The larger part of the barite mined in the United States is a residual mineral, forming concretions in clay resulting from the decay of limestone. It is sometimes difficult to separate the residual and the strictly epigenetic deposits for it appears that the same solutions which formed the concretions deposited barite in fracture zones in the underlying rock.

Deposits in the United States.—The barite deposits now worked are mostly contained in the Paleozoic limestones in the southern Appalachian and the central States, the order of importance being Missouri, Georgia, Tennessee, Kentucky and Virginia.

In Missouri barite often accompanies the zinc and lead deposits but the important deposits are found in a separate area in Washington County, in southeastern Missouri,[1] not far from the great lead mines in the Bonneterre (Cambrian) dolomite (p. 461). The principal deposits are found in the shattered and dolomitized Gasconade limestone (Ordovician) as filling of irregular veins and other open cavities. The order of precipitation is given by Steel as follows: A thin coating of chalcedony was first deposited; this was followed by deposition of a little galena; and this in turn was succeeded by barite, which is the main filling. The series of events was closed by the precipitation of marcasite, dolomitization, and the formation of a second generation of barite, and by a much later coating of ruby-red sphalerite on the older barite.

In Georgia barite occurs in the Cartersville district[2] as deposits from solution in fractures and cavities in the Weisner quartzite in intimate association with yellow ocher, and also as nodules embedded in residual clays.

The barite deposits of Virginia have been described by T. L. Watson,[3] who states that they are probably caused by the leaching of limestones by meteoric waters. Deep rock decay character-

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[1] A. A. Steel, *Trans.*, Am. Inst. Min. Eng., vol. 40, 1910, pp. 85–117.
Arthur Winslow, Missouri Geol. Survey, vol. 7, 1894, p. 678.
E. R. Buckley, Missouri Bureau of Geology and Mines, vol. 9, pt. 1, 1908.

[2] T. L. Watson and J. S. Grasty, Barite of the Appalachian States, *Trans.*, Am. Inst. Min. Eng., vol. 51, 1916, pp. 514–559.

[3] *Trans.*, Am. Inst. Min. Eng., vol. 38, 1907, pp. 953–976.
T. L. Watson and J. S. Grasty, *op. cit.*

izes the whole region. The barite in part fills fractures and in part replaces limestone. It occurs:

1. In crystalline Cambrian or pre-Cambrian limestone as irregular, lenticular lodes or pockets replacing the limestone and associated with calcite and chalcopyrite (Fig. 124).

2. In crystalline schists as filling of fractures.

3. In the Shenandoah (Cambro-Ordovician) limestone as filling of fractures or in residual soil.

4. In Triassic shales and limestone as filling of fissures in a crushed zone.

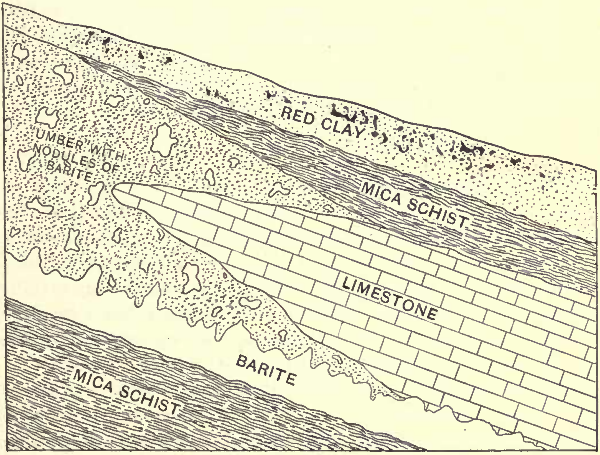

FIG. 124.—Section of the Bennett mine, Virginia, showing occurrence of barite as residual and as replacement deposit. *After T. L. Watson.*

Large and pure barite veins have been described from the Ketchikan district[1] and near Wrangell[2] in Alaska. They are contained in crystalline limestone and in schists.

Foreign Deposits.—Barite deposits are common in all countries. Bodies of exceptional size and purity are found in central Germany in sedimentary rocks of Permian and Triassic age.

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[1] T. Chapin and G. H. Canfield, *Bull.* 642, U. S. Geol. Survey, 1916.

[2] E. F. Burchard, *Bull.* 592, U. S. Geol. Survey, 1914, pp. 109–117.

Uses and Production.—Barite is used extensively as a pigment in the manufacture of mixed paint and to give weight to paper. It is the raw material for other barium salts, such as the nitrate, which is used in pyrotechnics for green fire. For most of the purposes indicated its purity and white color are essential. The crude material is crushed and treated in log washers and jigs. After grinding the pulp is classified and the settled cream-colored mud is finally treated with sulphuric acid to remove the staining ferric hydrate. The domestic production in 1917 was about 207,000 tons, which came from a great number of small operators in Missouri, Virginia, Kentucky, Georgia, North Carolina, and Tennessee. Before the European war began from 16,000 to 30,000 tons of barite and barium salts were imported annually, largely from England and Germany. The removal of competition with high-grade European barite has greatly stimulated the American industry. The average price was $5.66 per ton.

CELESTITE AND STRONTIANITE[1]

Strontium accompanies barium as a primary constituent of igneous rocks but is present in much smaller quantities. As celestite ($SrSO_4$) and strontianite ($SrCO_3$) it is sometimes found in fissure veins of hydrothermal origin, but the two minerals are much more commonly found as veins, nodules and layers in sedimentary rocks, particularly limestone. In the latter case they are undoubtedly leached by cool surface waters from small quantities in the sediments and deposited in convenient places.

Celestite is found in crystals and granular masses often of bluish color; but sometimes it is dark or brownish. It usually, but not always, contains $BaSO_4$. Strontianite is crystalline, fine-grained, fibrous or nodular and has white, brownish or dark color. It has often been mistaken for calcite, and always contains a few per cent. of $CaCO_3$.

In geods, veins, disseminations and replacements celestite is found in Paleozoic dolomite and limestone of Michigan,[2] New York[3] and Ohio. A cave at Put-in Bay is said to have yielded

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[1] J. M. Hill, Barytes and strontium, *Mineral Resources*, U. S. Geol. Survey, pt. 2, 1915, pp. 161–187.

[2] E. H. Kraus and W. F. Hunt, *Am. Jour. Sci.*, 4th ser., vol. 21, 1906, p. 237. W. A. Sherzer, *Am. Jour. Sci.*, 3d ser., vol. 50, 1895, p. 246; *Rept.*, Michigan Geol. Survey, vol. 7, pt. 1, 1900, p. 208.

[3] E. H. Kraus, *Am. Jour. Sci*. 4th ser., vol. 18, 1904, p. 30; vol. 19, 1905, p. 286.

150 tons of celestite. Deposits in the limestone quarries of northwestern Ohio and southeastern Michigan, near Toledo are said to be of possible economic importance. Strontianite is subordinate. Celestite also occurs in Cretaceous limestone in Texas[1] near Austin. Interesting deposits of celestite have lately been discovered in Tertiary lake beds in Arizona,[2] near Gila Bend where the mineral occurs in sandstone and shale with gypsum and salt; similar beds have been described from the Awavatz mountains near the southern end of Death Valley. Strontianite deposits in Miocene lake beds 10 miles north of Barstow,[3] San Bernardino County, have been discussed by A. Knopf who believes that the beds, veins and fibrous concretions at this place are replacements of lacustrine limestone.

All these deposits are of low grade but it is planned to utilize them at the present time. In 1916, 250 tons of strontium ore was mined in the United States, the first production for many years. In 1917 the output had increased to 4,035 short tons.

The principal supply of strontium was for many years derived from the strontianite veins in Cretaceous marl and limestone of Westphalia, Germany; the mineral is here accompanied by calcite and a little pyrite.[4]

The strontium used in the United States before the war came largely from England. Important celestite deposits are found near Bristol, where the mineral forms lenses and veins in a Triassic marl and in the underlying rocks.[5]

Celestite is also concentrated, like barite, during the formation of sulphur from gypsum and as noted below under "Sulphur" it occurs in considerable quantities in the sulphur mines of Sicily. For commercial purposes celestite should contain at least 95 per cent. $SrSO_4$.

The principal use of strontium is in sugar refining, in the so-called Scheibler process, in which strontium hydroxide is used

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[1] First report, Geol. Survey Texas, 1889, p. 125.

F. L. Hess, *Eng. and Min. Jour.*, July 17, 1909, p. 117.

[2] W. C. Phalen, Celestite deposits in California and Arizona, *Bull.* 540, U. S. Geol Survey, 1914, pp. 526–531.

[3] Strontianite deposits near Barstow, California, *Bull.* 660, U. S. Geol. Survey, 1918, pp. 257–270.

[4] Götting, *Oesterr. Zeitschr. B. u. H. Wesen*, vol. 37, 1889, p. 113.

[5] R. L. Sherlock, *Mem.*, Geol. Survey England, Special reports on mineral resources, vol. 3, 1918, pp. 48–61.

for the recovery of sugar from beet sugar molasses. The nitrate is used in pyrotechnics for red fire. The domestic production has been stimulated because of the recent embargo on exports from Great Britain. The price of British celestite was about $12 per ton at the eastern sea board.

SULPHUR[1]

Modes of Occurrence.—Native sulphur may be formed by various reactions. The oxidation of pyrite sometimes results in crusts of sulphur coating the cavities once occupied by the dissolved crystals. In the craters of volcanoes where sulphurous gases ascend on crevices sulphur is often found, as the result of a reaction between sulphur dioxide and hydrogen sulphide ($H_2S+2SO_2=H_2SO_4+2S$), or more probably by incomplete oxidation of hydrogen sulphide ($2H_2S+O_2=2H_2O+2S$) or by the by the reaction $3SO_2+2H_2O=2H_2SO_4+S$. A large deposit of this kind is worked at the Abosanobori mine, Hokkaido, Japan, and consists of clayey beds in an old crater lake. Considerable quantities are exported from Japan to the United States. It has been proposed to utilize a similar deposit in the crater of Popocatepetl, Mexico; other deposits are found in the volcanoes of the Chilean and Argentine Andes.

Much more commonly sulphur is found at active or extinct hot springs in the tufas or other adjoining porous rocks like volcanic tuffs. It evidently results from the incomplete oxidation of H_2S, by the oxygen or by bacterial action. Such deposits have been observed at many places in the Western States—for instance, at Cuprite, Esmeralda County, Nevada; at the Rabbit Hole mines in Humboldt County, Nevada;[2] at Sulphur Bank, California; at the Cove Creek mine, Beaver County, Utah;[3] and at Cody and Thermopolis, in Wyoming.[4] The three last-named deposits have been worked. In Wyoming the sulphur in part

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[1] O. Stutzer, Die wichtigsten Lagerstätten der Nicht-Erze, 1911, pp. 185–263.

W. C. Phalen, *Mineral Resources,* U. S. Geol. Survey, Annual publication.

S. H. Salisbury, *Mineral Industry,* Annual publication.

[2] G. I. Adams, *Bull.* 225, U. S. Geol. Survey, 1904, pp. 497–500.

[3] W. T. Lee, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 485–489.

[4] E. G. Woodruff, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 451–456.

E. G. Woodruff, *Bull.* 380, U. S. Geol. Survey, 1909, pp. 373–380.

replaces the limestone underlying the travertine or tufa. All these deposits are superficial, and, though some are fairly productive, they play no great part in the world's output.

The greater part of the native sulphur is not connected with volcanic processes or hot springs but is obtained from sedimentary beds, in close association with gypsum and limestone; calcite, aragonite, barite, celestite, opal, more rarely quartz, together with gaseous and solid hydrocarbons, are found with the sulphur. This association is constant and characteristic and recurs in almost all the great gypsum beds of the world, though the sulphur is not always present in quantities of economic importance. As an illustration it is interesting to scan the boring records in Louisiana, contained in the bulletins of the State Survey, and note the frequency with which sulphur accompanies gypsum. The sulphur is in earthy or resinous masses and forms lenticular beds, veinlets, and concretions in marl, limestone, and gypsum.

FIG. 125.—Banded sulphur rock from Sicily, one-half original size. Black, sulphur; white, calcite; stippled, limestone. *After O. Stutzer.*

Origin of Sulphur Deposits in Gypsum.—Sulphur is undoubtedly derived from gypsum through the reducing action of organic matter, by way of calcium sulphide and hydrogen sulphide. Regarding the details of the transformation the views are not uniform; it is certain that the reaction can take place at low temperature. G. Bischof, in the middle of the last century, first discussed this matter[1] and assumed the following reactions:

$$CaSO_4 + 2C = CaS + 2CO_2.$$
$$CaS + CO_2 + H_2O = CaCO_3 + H_2S.$$
$$H_2S + O = H_2O + S.$$

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[1] G. Bischof, Chemische und physikalische Geologie, vol. 2, 1851, pp. 144–164.

The objection to this scheme would be that the sulphur is evidently often formed at depths of several thousand feet, and that the presence of much oxygen at such depths would be improbable; more likely the hydrogen sulphide generated from the gypsum reacts upon calcium carbonate, resulting in secondary gypsum and sulphur.

The deposits of Sicily have been the subject of extended discussion. A. von Lasaulx[1] has regarded them as formed in fresh-water lakes into which springs containing H_2S were discharged. G. Spezia[2] has advanced a similar view, believing, however, that the hot springs deposited the sulphur at the bottom of a sea basin, accounting for the presence of celestite by the same agency.

Baldacci[3] held that the deposition of sulphur took place in a partially evaporated marine basin, in or near which numerous mud-volcanoes, like those of the Apsheron peninsula in the Caspian Sea, discharged large volumes of hydrocarbon that effected the reduction of gypsum to calcium sulphide.

A theory of the purely sedimentary origin of the sulphur deposits of Sicily was recently advanced by O. Stutzer. The well-defined stratification of the sulphur beds, with occasional cross-bedding, the occurrence of the sulphur in limestone and its absence in the overlying gypsum, and finally the presence of intercalated clay beds which would prohibit the free circulation of water are cited by Stutzer as proofs of his view. Sedimentary sulphur deposits may form, according to him, in any closed basin in which hydrogen sulphide is developed. The gas may be produced by decay of organisms, or by reduction of dissolved calcium sulphate by carbon, or by hydrocarbons. The oxidation of hydrogen sulphide is effected by the oxygen of the air or by the aid of bacteria. In organic decay many bacteria reduce sulphates and develop hydrogen sulphide. Other low organic forms, the so-called sulphur bacteria, oxidize H_2S and accumulate sulphur in their cells as minute particles. The oxidation of this sulphur supports the life of the organism, the resulting sulphuric acid being converted into sulphates by

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[1] *Neues Jahrbuch*, 1879, pp. 490–517.

[2] G. Spezia, Sull' origine del solfo nei giacimenti solfiferi della Sicilia, Torino, 1892. Reviewed in *Neues Jahrbuch*, 1893, 1, p. 281.

[3] Descrizione geol. dell Isola di Sicilia, *Mem. descritt. d. Carta geol. d'Italia*, 1, 1886.

carbonates which are absorbed and are necessary for the growth of the bacteria. The sulphur bacteria are found in sulphur springs and in the mud of seas and lakes, in which hydrogen sulphide is developed. Stutzer also refers to the fact that the water of closed basins, such as the Black Sea, contains H_2S in quantities increasing with the depth. Stutzer's conclusions are supported by W. F. Hunt[1] who describes the bacterial action in detail.

Interesting as these views are, the sedimentary deposition of sulphur cannot be regarded as proved. The presence of epigenetic sulphur throughout large masses of gypsum is so common that its origin through the direct reduction of gypsum, by way of hydrogen sulphide and organic matter, can scarcely be doubted. Conditions in the Black Sea would seem to be favorable for the deposition of sulphur and yet no sulphur appears to have been brought up by deep dredgings in that basin.[2]

Examples.—Sulphur is widely distributed in the Miocene and Pliocene of the Mediterranean countries, everywhere accompanied by gypsum. By far the most important deposits are in Sicily, which for years has supplied the bulk of the world's production.

The sedimentary rocks of Sicily, in part marine, in part land deposits, consist of basal clays covered by diatomaceous and radiolarian shales. Above these beds the sulphur-bearing gypsum formation extends over an area of almost 800 square kilometers. This formation is about 300 feet in thickness; gypsum, limestone, salt, clay, and sandstone are the principal rocks. There are three or four beds of sulphur, the substance ramifying through the bluish-gray limestone. Celestite occurs in economically important quantities and with sulphur, gypsum, calcite, and more rarely barite forms beautiful crystals coating the walls of cavities. The crude ores of Sicily contain from 8 to 25 per cent. sulphur.

As noted above, sulphur is common in the Tertiary and Cre-

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[1] The origin of the sulphur deposits of Sicily. *Econ. Geol.*, vol. 10, 1915, pp. 543–579.

[2] Sir John Murray, The deposits of the Black Sea, Scottish Geog. Mag., 16, 1900, pp. 673–702. Stelzner and Bergeat, Erzlagerstätten, p. 470. A comprehensive review by Doss is given in the *Neues Jahrbuch*, 1900, 1, pp. 224–228.

taceous beds underlying the Louisiana and Texas coast.[1] In 1865 an unusually large deposit was discovered in Calcasieu Parish (230 miles west of New Orleans), Louisiana, at a depth of 443 feet underneath clay, sand, and limestone of Tertiary and Cretaceous age. The borings showed a thickness of 100 feet of almost pure sulphur, underlain by a great thickness of sulphur-bearing gypsum (Fig. 126). The lateral extent is sharply defined as a circular area, half a mile in diameter. It is now evident that these deposits have been formed in the upper part of one of the great salt domes of the Gulf coast (Fig. 102, see also p. 309). The difficulties of sinking a shaft through the quicksands for a long time prevented the utilization of this deposit. Later, however, the difficulties were overcome by the invention of the Frasch process.[2] Through bore holes superheated water is forced down to the sulphur, which is thereby melted; hot air is then supplied under pressure to aerate the molten mass and facilitate its ascent by water pressure to the surface

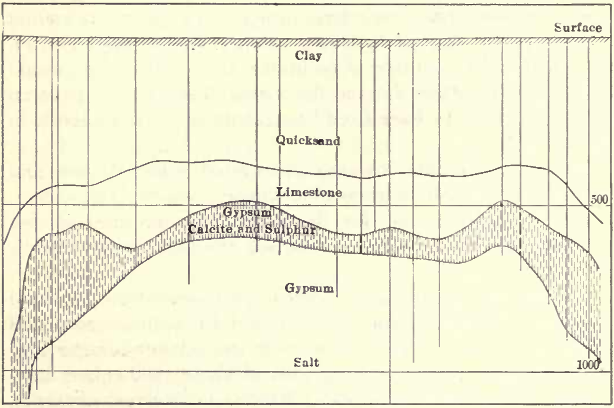

FIG. 126.—Vertical section of sulphur-bearing bed at Calcasieu parish, Louisiana. *After Kirby Thomas.*

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[1] J. F. Kemp, *Mineral Industry*, 1893, p. 585.
W. C. Phalen, *Mineral Resources*, U. S. Geol. Survey, 1907, pt. 2, p. 674. See also issue for 1911, pt. 2.

[2] H. Frasch, *The Mining World*, December 14, 1907.
W. C. Phalen, *Op. cit.*

Similar sulphur beds occur at several places along the Gulf Coast. A deposit has recently been opened by drilling by the Freeport Sulphur Company near the mouth of the Brazos River, Texas, and is now in successful operation. Here the sulphur beds lie under 760 feet of gravel, sand and clay. Below this are 150 feet of sulphur bearing limestone, gypsum and dolomite, containing from 10 to 50 per cent. sulphur. These beds are underlain by gypsum, limestone, sandstone and rock salt.[1] For some reason the literature on Texas and Louisiana sulphur deposits is extremely scant.

Production.—Until recently the Sicilian deposits, with an annual output of about 450,000 metric tons, supplied the world's demand. In 1901 the Frasch process revolutionized the trade conditions and the production of the United States rose at once to 200,000 or 300,000 long tons, and the importation from Sicily fell off correspondingly. The interesting trade conditions developing from these changes and the struggles of the Italian government to aid the distressed Sicilian operators are described in the articles in *Mineral Resources* cited above.

In 1915 the Italian production was 364, 260 metric tons while that of the United States was in excess of 400,000 tons with rapidly diminishing imports. The price was about $17 per ton. In 1917 the sulphur production of the United States was probably much larger though, on account of the output being divided between two principal producers, the figures are no longer published by the U. S. Geological Survey. Louisiana, Texas, Wyoming and Nevada are the producing states.

Uses.—The manifold industrial uses of sulphur need not be specified; the larger part is used for the manufacture of sulphuric acid, for bleaching purposes by the development of sulphur dioxide, for the prevention of mildew on grapevines, and for the manufacture of gunpowder, matches, etc.

Sulphuric Acid.—Just before the war the production of sulphuric acid in the United States was about 3,500,000 short tons. War conditions soon created an enormous demand for sulphuric acid, mainly for explosives and the price has risen rapidly. In 1916 about 6,250,000 tons of acid (50° Baumé) was produced and for 1918 the requirements were 8,000,000 tons. Of the production in 1916, 40 per cent. was made from

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[1] W. C. Phalen, quoting Thomas Kirby in *Mineral Resources*, U. S. Geol. Survey, 1912, p. 936.

Spanish (Rio Tinto) pyrite, 6 per cent. from Canadian nyrite, 13 per cent. from domestic pyrite, marcasite and pyrrhotite, 22 per cent. from fumes from copper and zinc smelters, leaving about 19 per cent. which had to be supplied from native sulphur[1] which practically did not appear at all in the acid production of normal times. The scarcity of shipping has reduced the importations from Spain so that great efforts have been made to develop our native supplies of pyrite and sulphur.

It appears then that pyritic ores with or without other metals are the principal source of sulphuric acid. Many countries, particularly Spain, Norway, Portugal, France, United States, Italy and Germany in the order of importance stated, produce annually over 200,000 tons of pyrite.

In the United States pyrite, marcasite and pyrrhotite with, respectively, 53.3 and 38.4 per cent. of sulphur are the principal sulphur ores. They are obtained:

1. From pyrite deposits along the Appalachian mountains from Alabama to Vermont.
2. From pyrite deposits in California.
3. From pyrrhotite deposits in Virginia, Tennessee and Maine,
4. From marcasite as a by-product of coal mines in Illinois, Ohio, Indiana and Pennsylvania.
5. From marcasite as a by-product in zinc-lead mines of Wisconsin and Illinois.

The domestic production of pyrite (including marcasite and pyrrhotite, was about 400,000 long tons of which the larger part came from Virginia and California. Pyrite was also imported from Quebec and Ontario.

The "pyritic deposits" comprise many types (p. 635), but aside from the minor supplies mentioned under 4 and 5, they are mainly products of high or intermediate temperature under intrusive conditions and most of them may be considered as copper deposits of very low grade. Many among those along the Appalachian belt are of early Paleozoic age and more or less strongly dynamo metamorphosed. We may mention the pyrrhotite deposits of Ducktown, Tennessee, and of the "Great Gossan lead," Virginia, further the pyritic deposits of Louisa County, Virginia, which form long lenses in a Cambrian sericite schist and northward change into lead and zinc deposits. Other

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[1] W. Y. Westervelt and A. G. White, *Bull.* 130, *Trans.*, Am. Inst. Min. Eng., October, 1917, pp. 5–15.

deposits of pre-Cambrian age are found in St. Lawrence County, New York. The ores contain from 30 to 50 per cent. of sulphur with up to 10 per cent. of insoluble. Lead, zinc, antimony and arsenic are objectionable constituents. The residues from sulphuric acid manufacture are often used as copper and iron ores.

THE MAGNESIAN DEPOSITS

The magnesian silicate rocks lend themselves easily to transformation and yield a number of economically valuable products, among them serpentine, magnesite, meerschaum, talc, soapstone, and asbestos. All of these result from the action of water, in most cases doubtless of atmospheric origin, on peridotites, pyroxenites, or gabbros, either near the surface or with the co-operation of stress at greater depths. Talc, soapstone, and asbestos belong, in part, to the latter class.

SERPENTINE[1]

Serpentine forms by simple hydration from a rock consisting of enstatite and olivine according to the following equation:

$$\underset{\text{(Olivine)}}{Mg_2SiO_4} + \underset{\text{(Enstatite)}}{MgSiO_3} + 2H_2O = \underset{\text{(Serpentine)}}{H_4Mg_3Si_2O_9}.$$

It may also develop from olivine alone, with the removal of some magnesium as carbonate:

$$2Mg_2SiO_4 + CO_2 + 2H_2O = H_4Mg_3Si_2O_9 + MgCO_3.$$

The latter equation probably represents the usual process of serpentinization a short distance below the surface. Under oxidizing conditions serpentine is unstable, though of course the change takes place very slowly and erosion may work far ahead of decomposition.

Serpentine is, however, also formed on a large scale at greater depths, where quantities of CO_2 could not very well be assumed for the reason that such alteration would result in a mixture of serpentine and carbonates, whereas the large serpentine masses rarely contain admixed carbonates. The deep cañons of the Sierra Nevada, in California, show clearly that the serpentines of this range are not superficial, but descend to the depth of several thousand feet. The *modus operandi* of such exten-

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[1] H. Leitmeier in Doelter's Mineralchemie, vol. 1, pt. 1, 1914, pp. 385–428.

sive hydration is not fully explained. Some have held that it might have been effected by ascending waters, shortly after the intrusions.

Serpentine is generally rich in iron, for the original rocks are not of the purity indicated by the equation given above; the iron is present both as silicate and magnetite, and also in the chromite which forms a characteristic accessory. Rock that is not too much broken by joints finds fairly extensive use as building and ornamental stone. For the latter purpose the oily green translucent varieties formed in crystalline limestone by serpentinization of the contained pyroxene are particularly valued.

MAGNESITE[1]

Origin.—Magnesite ($MgCO_3$) appears in two modifications: (1) As an amorphous, earthy, hard and compact mineral, which probably is a hardened colloid precipitate. It is often concretionary and has a conchoidal fracture like that of unglazed porcelain. In this form it is an alteration product of serpentine or allied magnesian rocks as illustrated by the equation-$H_4Mg_3Si_2O_9+3CO_3=3MgCO_3+2H_2O+2SiO_2$. (2) As a crystalline mineral, isomorphous with calcite and usually holocrystalline granular. In this form it is generally a replacement of dolomite produced by magnesian solutions in connection with intrusions.[2]

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[1] Robert Scheerer, Der magnesit, Vienna und Leipzig, 1908, p. 256.

M. Dittrich, H. Leitmeier, K. A. Redlich in Doelter's Handbuch der Mineralchemie, Dresden and Leipzig, 1912, vol. 1, pp. 212–267.

F. L. Hess, The magnesite deposits of California, *Bull.* 335, U. S. Geol. Survey, 1908.

C. G. Yale and H. S. Gale, *Mineral Resources*, U. S. Geol. Survey, Annual publication.

S. H. Dolbear, *Mineral Industry*, Annual publication.

[2] Certain minor occurrences are of interest: Magnesite of the amorphous type is found as sedimentary beds and lenses in clays of Miocene lake beds near Bissell, San Bernardino County, California. See H. S. Gale, *Bull.* 540, U. S. Geol. Survey, 1914, p. 512.

Crystalline magnesite occurs in many crystalline schists of the Austrian Alps.

Hydromagnesite ($3MgCO_3.Mg(OH)_2 + 3H_2O$) is reported from Atlin, British Columbia, as a deposit of fine white powder several feet deep and appearing like a spring deposit. In connection with this it is recalled that H. Leitmeier found that a magnesian hydrocarbonate was deposited by the mineral waters of Rohitch in Styria. *Zeitschr. Kryst. Min.*, vol. 47, 1909, p. 118.

Occurrence.—The amorphous magnesite is not uncommon in areas of serpentine, and it occurs in fissures or crush-zones or irregular masses, often mixed with more or less serpentine and some opal or chalcedony. It is often very pure with slight admixtures of iron, alumina and lime; a few per cent. of free silica are often present. Magnesite occurs abundantly but generally in small deposits in the California[1] Coast Ranges; the best deposits are near Porterville, Tulare County. Until recently the production has been small owing to distance from the eastern market. Similar deposits of great extent producing annually 134,000 tons are worked on the coast of Euboea, in Greece. Other localities are found at Salem, near Madras, India, in the Transvaal, and many other lands where serpentinoid rocks abound. Magnesite does not always accompany serpentine, however, and it may be surmised that ascending springs with much CO_2, as are so common in California, may have some connection with its genesis.

The largest magnesite deposits in the world are at Veitsch, in Styria,[2] Austria, where in 1914 about 200,000 metric tons were mined in open quarries. This crystalline magnesite is a replacement of dolomite formed under the influences of intrusions of porphyry and other acidic and basic rocks. The Austrian magnesite, though otherwise pure, contains like that from Greece a few per cent. of iron which makes it desirable for basic linings and bricks. In 1914, 110,000 tons of this material were imported and the war created a shortage which stimulated local production and search. In 1916 a large magnesite deposit was discovered near Chewelah, Washington, which bids fair to supply the demand. Like the Styrian deposits it is a replacement of dolomite of Carboniferous age, near granite; in places the material contains a few per cent. of silica, lime and iron. Very similar are also the deposits found some years ago in the Grenville township,[3] Quebec, and which are

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[1] F. L. Hess, *Op. cit.*

H. S. Gale, *Bull.* 540, U. S. Geol. Survey, 1914, pp. 483–520.

[2] K. A. Redlich, Die Genese der Pinolitmagnesite, Siderite und Ankerite der Ostalpen, Tsch. m. und petr. Mitt., vol. 26, 1907, pp. 499–505.

K. A. Redlich, Genesis der kristallinen Magnesite, *Zeitschr. prakt. Geol.*, vol. 21, 1913, pp. 90–101.

[3] H. J. Roast, The development of Canadian magnesite, *Trans.*, Canad. Min. Inst., vol. 20, 1917, pp. 237–255.

M. E. Wilson, Magnesite deposits of Grenville districts, etc. *Mem.* 98, Canada Geol. Survey, 1917.

now being worked. At this locality the magnesite contains several per cent. of lime and very little iron which to some degree has made difficult its use for refractories.

Production and Use.—The domestic production advanced sharply in 1916 to 154,000 tons, and in 1917 reached 316,000 tons. The price was about $10 per ton. Magnesite gives off its carbon dioxide at 800° C., and is, therefore, preferred to calcite in the production of this gas. After calcining, the substance is used for the manufacture of various magnesium salts, and in the paper and sugar industries. It is employed extensively with magnesium chloride for the so-called Sorel cement, used for flooring, etc. Its most important use is for basic furnace lining in the Thomas process. Until recently the Styrian magnesite was imported mainly for this purpose. Magnesite for bricks should contain a few per cent. of FeO and little CaO; 8 per cent. CaO, being the allowable limit.

Magnesite serves also as an ore for the production of metallic magnesium, which on account of its low specific gravity (1.74) is now used in alloys with aluminum and other metals. The reduction is effected by treating the chloride made from magnesite in an electric furnace. The best ore for the purpose is naturally the carnallite ($KCl.MgCl_2+6H_2O$) from the Stassfurt salt beds (p. 312). The production of magnesium in United States in 1917 was 116,000 pounds, the price being about $2.00 per pound.

MEERSCHAUM[1]

Meerschaum or sepiolite ($H_4Mg_2Si_3O_{10}$, containing SiO_2, 60.8 per cent.; MgO, 27.1 per cent.; H_2O, 12.1 per cent.) is a hydrated silicate of magnesia of tough, compact texture, white or cream color, and smooth feel. As is well known, it finds a rather extensive use in the manufacture of pipes and cigar holders. Its analysis usually shows a little iron, alumina, and lime. It is probably derived from serpentine by slow hydration and is in most cases a colloid precipitate. The principal occurrence is in Asia Minor at Eski-Shehr, where it is found as nodular masses near the surface; at this and several other localities in Crimea and Bosnia serpentine rocks are found in the vicinity, although the material itself is embedded in Quaternary or Tertiary beds.

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[1] G. P. Merrill, Non-metallic minerals, 1910, pp. 218–221.
C. Doelter, Mineralchemie, vol. 2, pt. 1, 1914, pp. 374–383.

A different occurrence is that recently discovered in New Mexico,[1] on the upper Gila River, where the substance forms veins and balls in a Paleozoic cherty limestone. Here it is probably derived from a dolomitic carbonate.

TALC AND SOAPSTONE[2]

General Occurrence and Origin.—Talc ($H_2Mg_3Si_4O_{12}$, or $3MgO.4SiO_2.H_2O$; 65.5 per cent. SiO_2, 31.7 per cent. MgO, 4.8 per cent. H_2O), is a hydrated magnesium silicate, but holds much less water than serpentine. It is a soft, crystalline, foliated or compact mineral of white, gray, or pale-green color and a greasy feel. The more compact, as well as some impure varieties, are usually called soapstone; they may contain shreds of chlorite and other ferromagnesian minerals, like enstatite or amphibole. Soapstone is easily worked and is of great resistance to acids and high temperatures.

Talc and soapstone are products of the hydration of magnesian rocks, either of distinctly igneous origin, like gabbro, pyroxenite, or peridotite, or crystalline schists rich in such minerals as enstatite and tremolite or other pyroxenes and amphiboles. These schists may result from the shearing of igneous or contact-metamorphic rocks, the latter derived from the igneous alteration of limestone and dolomite. The purest talc deposits are associated with crystalline carbonate rocks containing amphibole. In general serpentine forms from olivine and talc from pyroxene and amphibole, but this rule does not always hold.

Talc often contains 1 or 2 per cent. of iron and aluminum, as well as a little calcium; according to the analyses given by Merrill (*op. cit.*) the soapstones contain, in addition to silica and magnesia, from 5 to 11 per cent. alumina, 7 to 13 per cent. ferrous oxide, and 1 to 4 per cent. lime; some of them contain so much water that a strong admixture of serpentine must be assumed.

The formulas show that talc may be obtained from enstatite or tremolite by the addition of water and carbon dioxide, with separation of magnesium or calcium carbonate, which is probably carried away in solution; or, in case of deficiency of CO_2, the

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[1] D. B. Sterrett, *Bull.* 340, U. S. Geol. Survey, 1908.

[2] J. S. Diller, *Mineral Resources*, U. S. Geol. Survey, Annual issues.
G. P. Merrill, *op. cit.*, pp. 208–216.
C. Doelter, *op. cit.*, pp. 356–374.

magnesia may combine with silica, possibly set free from other minerals, to form additional talcose material.

The exact conditions and temperature needed for the formation of talc are not known, but it seems certain that dynamic stress, together with a limited supply of water not over rich in CO_2, is favorable to its development; it also undoubtedly forms from magnesian minerals by the aid of a scant supply of surface water under static conditions. It is also known that talc may develop along fissures under the influence of ascending hot waters, whenever magnesian silicate rocks are traversed.

E. Weinschenk,[1] in his description of the talc deposits of the Austrian Alps, holds that the mineral develops by replacement of schist composed of quartz, chlorite, chloritoid, and graphite along its contact with limestone and believes this transformation due to waters following the irruption of large igneous bodies.

Occurrences.—The crystalline schists of all countries yield talc. Some occurrences are known from the Pacific coast, but the production in the United States is limited exclusively to the belt of ancient crystalline rocks which forms the axis of the Appalachian Mountain system from Canada to Alabama.

North Carolina is rich in talc, and one belt of Cambrian marble along the Nantahala Valley and Nottely River[2] yields many lenses as much as 200 feet long and 50 feet thick. The mineral is mined in open cuts and by shafts and tunnels.

New York and Vermont easily outrank all other States in the production of talc. The output of New York comes from a small district about 12 miles southeast of Gouverneur,[3] which has been worked for many years by underground methods. One mine at Talcville has attained a depth of 550 feet. The mineral occurs in schistose layers of enstatite and tremolite, gradually merging into the surrounding crystalline limestone. The deposit forms a persistent layer averaging 20 feet in width, within the enstatite-tremolite rock.

Virginia yields most of the soapstone produced in the United

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[1] *Abhandl.* Bayer. Akad. d. Wiss., vol. 21, pt. 2, 1901, p. 270.

[2] Arthur Keith, *Bull.* 213, U. S. Geol. Survey, 1903, p. 443.

J. H. Pratt, North Carolina Geol. Survey, *Economic Paper* 3, 1900, p. 99.

J. H. Pratt, *Mineral Resources*, U. S. Geol. Survey, 1905, p. 1361.

[3] C. H. Smyth, Jr., *School of Mines Quarterly*, vol. 17, 1896, pp. 333–341.

States. It is derived from a belt nearly 30 miles long and less than one mile wide. According to T. L. Watson's description[1] the soapstone occurs as sheets or dike-like masses, 100 feet or more in thickness, conformably interbedded with quartzitic schists, but is probably derived from an igneous rock.

Production and Uses.—The rapidly expanding production of talc in the United States was about 198,600 tons in 1917. The larger part was sold in powdered form. The value of powdered talc is about $10 per ton.

Talc is used as a filler for paper, including wall paper; also for admixture or adulteration of pigment, as a heat insulator, lubricant, polishing powder of glass, for toilet powders, and as an absorbent for nitroglycerine. The compact talc or soapstone is used for fire-bricks, laboratory tables, gas burners, crayons, etc.

Pyrophyllite.[2]—Pyrophyllite is a hydrous silicate of alumina ($H_2Al_3Si_4O_{12}$), containing 66.7 per cent. SiO_2, 28.3 per cent. Al_2O_3, and 5.0 per cent. H_2O. In composition and physical qualities it is similar to talc, though it does not command as high a price as the best talc. It is mined in Moore and Chatham counties, North Carolina, where it occurs in thick beds associated with slate.

ASBESTOS[3]

Amphibole Asbestos.—The asbestos of mineralogy is a monoclinic amphibole which develops in seams and slips in normal amphibolitic rocks, especially where the rocks have been subjected to pressure and movement. Chemically it is a calcium-magnesium metasilicate. According to the series of analyses given by Merrill the silica varies from 52 to 58 per cent., the lime from 12 to 16 per cent., the magnesia from 20 to 30 per cent. Other constituents are alumina, varying from 1 to 6 per cent., and ferrous oxide, usually from 1 to 6 per cent., though in some cases considerably higher. Water is always present, the amount generally varying between 2 and 5 per cent. Although contrary to the views of some authorities, the conclusion can

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[1] T. L. Watson, Mineral Resources of Virginia, 1907, p. 293.

[2] J. H. Pratt, *Op. cit.*

[3] G. P. Merrill, *Proc.* U. S. Nat. Mus., vol. 18, 1895, p. 181. *Bull.* Geol. Soc. Am., vol. 16, 1905, p. 113. Non-metallic minerals, 1910, pp. 183–197.

F. Cirkel, Chrysotile-asbestos, Canada Dept. of Mines, Mines Branch, 1910; 316 pp.

J. S. Diller, *Mineral Resources*, U. S. Geol. Survey, Annual publication.

hardly be avoided that the water is an essential constituent and that the mineral is really a hydrated form of tremolite or actinolite. The extinction angle appears, however, to be that characteristic of these amphiboles, or about 18°. No experiments appear to have been made as to the temperatures at which the water is driven off. The normal varieties of amphibole also hold a little water, but in far smaller quantities than asbestos.

Anthophyllite (Mg,Fe) SiO_3, and crocidolite, $NaFeSi_2O_6$.-$FeSiO_3$, a dark blue sodium amphibole, also yield asbestiform varieties.

Merrill has shown that the fibers are polygonal in outline and run out into needle-like points; down to a diameter of 0.002 or 0.001 millimeter the fibers retain their uniform diameter and polygonal outlines. The color of amphibole asbestos is usually white to greenish white. Only the finer kinds are utilized, but even these are less valued than the serpentine asbestos. They are apt to be less flexible and somewhat brittle.

Most of the small quantity of asbestos mined in the United States is of the tremolite or actinolite variety, and it often occurs in limestones which have been partly metamorphosed to amphibolitic rocks. The mineral is classed as slip-fiber or cross-fiber, according to the position of the fibers in the veinlets. The radial or divergent structures are designated as mass-fiber.

There are many occurrences, mainly in pre-Cambrian rocks along the Appalachian Mountain system, from Vermont to Alabama. One of the most important localities worked is at Sall Mountain, Georgia, where, according to Diller, the asbestos occurs in large lenticular masses in gneiss and is believed to be an altered igneous rock. Almost the entire domestic production is derived from Georgia.

Serpentine Asbestos (Chrysotile).—Chrysotile asbestos is green or yellowish-green and is easily reduced to a white fluffy state. The fiber is short, but of very uniform diameter and great divisibility and flexibility; the decomposing effect of hydrochloric acid also distinguishes it from amphibole asbestos. In composition it is practically identical with the purer kinds of serpentine. A typical analysis of the Canadian material yielded per cents. as follows: 42 SiO_2, 42 MgO, 14 H_2O, 1 FeO, and 1.7 Al_2O_3. Fig. 127 shows the appearance of the two kinds of asbestos.

This variety is found as veinlets, rarely over 6 inches thick, in serpentine or peridotite, and has almost always a cross-fiber—that is, the silky fibers lie perpendicularly to the plane of the veinlet.

The pure yellowish-green serpentine which occurs in contact metamorphic limestone and which is an alteration product of diopside sometimes contains chrysotile of exceptionally high grade. A deposit of such material is now worked in Arizona, northeast of Globe.[1]

FIG. 127.—Chrysotile (a) and amphibole (b) asbestos. *Photograph by J. S. Diller.*

Chrysotile veinlets may be found in almost any serpentine area, but they are rarely so abundant and large as to be of economic importance. The views regarding their origin differ.

Dresser shows that serpentinization in the Canadian deposits proceeded along irregular cracks in the peridotite, and the chrysotile veinlets are found in the center of the serpentinized bands.

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[1] J. S. Diller, *Mineral Resources*, U. S. Geol. Survey, 1917, pt. 2, p. 197.

These veinlets were interpreted by Pratt and Merrill as fillings of contraction cracks, but other authors are probably correct in considering them the result of a recrystallization of the serpentine, proceeding inward from the cracks.

S. Taber[1] believes that all cross-fiber veins are formed by a process of lateral secretion, the growing veins pushing aside the enclosing walls. Since, however, the material in the veins is derived from the serpentine itself it is not apparent why there is any need of increase of volume.

Until about 1895 the small quantity of asbestos used in the United States came from Italy. Since that date the development of the asbestos industry in Canada has been extremely rapid, and the Canadian mines now supply this country. The Canadian deposits[2] center in Asbestos Hill at Thetford, in the eastern townships of Quebec. As stated, the mineral occurs as irregular veinlets in serpentine and peridotite. These rocks are in places accompanied by somewhat later gabbro and granite and all of them are intrusive into Ordovician sediments. The mineral is mined in open pits, one of which, for instance, is 700 feet long, 200 feet wide, and 165 feet in greatest depth. A small percentage is obtained by hand cobbing, but the larger part—30 to 60 per cent.—of the crude material quarried is crushed and screened, and the fibers are separated by air currents.[3] The extraction of fiber of the milled rock is from 6 to 10 per cent.

Of late years the Russian chrysotile from the Ural Mountains and the deposits in southern Rhodesia as well as the crocidolite asbestos from Griqualand West, Cape Colony, are becoming important. The large deposits of crocidolite occur in thin layers interbedded with jaspers and iron stones of the Pretoria series.

In the United States[4] chrysotile of economic importance is worked in Vermont, near Casper, Wyoming, and in Arizona. Thus far, the production is small.

Uses.—"The fundamental property of asbestos, upon which its use depends, is its flexible, fibrous structure, but coupled with this are the scarcely less important qualities of incombustibility

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[1] *Bull.* 120, Am. Inst. Min. Eng., Nov., 1916, pp. 1973–1998.

[2] J. A. Dresser, *Econ. Geol.*, vol. 4, 1909, pp. 130–140.

J. A. Dresser, Preliminary report on the serpentine, etc., of southern Quebec, *Mem.* 22, Canada Geol. Survey, 1913, p. 103.

[3] W. J. Woolsey, *Jour.* Can. Min. Inst., vol. 13, 1910, pp. 408–413.

[4] J. S. Diller, *Bull.* 470, U. S. Geol. Survey, 1910, pp. 505–524.

and slow conduction of heat and electricity when the mass is fiberized and porous." The spinning and weaving of fire-proof cloth form an important part of the asbestos industry carried on in the United States with Canadian raw material. The highest grade of the crude mineral is expensive, costing $275 to $350 per ton; the fines cost $25 to $125 per ton, while the lowest grade—a mixture of serpentine and asbestos—is sold at less than $1 per ton. Amphibole asbestos is much cheaper, costing about $18 per ton. Crocidolite is more easily fusible but is more resistant than chrysotile to acids and sea water. The London price is about $125 per ton.[1]

The Canadian production in 1917 was 144,185 tons; almost the whole production was exported to the United States. In the same year the output in the United States was 1,683 tons.

ORES OF COPPER, LEAD, VANADIUM, AND URANIUM IN SANDSTONE AND SHALE

General Features.—Ores of copper, lead, vanadium, and uranium are often found disseminated in sandstones and shales far from igneous rocks. The sedimentary strata containing the ores are usually parts of thick series of terrigenous or shallow-water beds, commonly of reddish color. The ores are of low tenor and can be utilized only in exceptional cases. Nevertheless this class of deposits presents many interesting features.

The ore minerals are chalcocite, galena, roscoelite (a vanadium mica), various copper and lead vanadates, carnotite (a vanadate of uranium), etc. Bornite, chalcopyrite, and pyrite are less common. The ores frequently carry small amounts of silver, nickel, cobalt, molybdenum, and selenium. Gangue minerals occur sparingly and are usually confined to a little barite, calcite, and gypsum. The outcrops are likely to be brilliantly colored by malachite and azurite. While the deposits are confined to certain formations or members, they do not continuously follow a particular horizon and give no evidence of being of sedimentary origin. They often appear in fractured and brecciated beds or in strata rich in carbonaceous matter and plant remains. More rarely the ores follow distinct fissures in the sedimentary rocks. They do not seem to have any genetic relation with thermal springs. The copper, lead, and vanadium

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[1] P. A. Wagner, South African Journal of Industries, Nov., 1917.

deposits form three groups in this class, but each group is likely to contain more or less of the other metals. There is no reason why the deposits should be confined to any particular geological age, but as a matter of fact almost all of them are in the upper Carboniferous, Permian, Triassic, or Jurassic.

Origin.—In considering the class as a whole it appears that igneous agencies had no part in the genesis. The ores are assuredly epigenetic and their universal appearance in land or shallow-water beds is significant. In all probability these ores have been concentrated by meteoric waters which leached the small quantities of metals disseminated in the strata. The sediments were rapidly accumulated, under arid conditions, from adjacent land areas and the metals were probably carried down in fine detritus and in solutions from older ore deposits in these continental areas.

The waters which concentrated the ores are believed to have been mainly sodium chloride and calcium sulphate solutions containing sulphates and perhaps chlorides of copper and lead. The mineral association and geological features indicate deposition at low temperature, probably well below 100° C., and at shallow depths but below the zone of direct oxidation. Very likely these ores have been forming continuously since the establishment of active water circulation in the beds; in favorable places below the surface concentration may now be in progress.

COPPER AND LEAD DEPOSITS IN SANDSTONE

European Occurrences.[1]—The European occurrences are confined to the Permian and the Triassic, both, generally speaking, ages of arid climate and saline deposits.

The Russian Permian, extending far west from the Urals, consists in its lower division of sandstones, marls (in part marine), and conglomerates. The sandstones are rich in vegetable remains. Copper ores are found over wide areas, but have not been worked extensively of late. The average tenor is said to be 0.9 per cent. metallic copper. The chalcocite ores replace plant remains and tree trunks or form the cement of the sandstones. The minerals mentioned from this locality are (besides

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[1] For an excellent review of European localities, as well as complete index of literature, in part difficultly accessible, see Stelzner and Bergeat, Die Erzlagerstätten, 1904, pp. 388–439.

secondary malachite and azurite) chalcocite, chalcopyrite, barite, vanadinite, and volborthite (vanadate of copper and calcium.)

Recently much interest has been taken in the copper deposits of the Khirgiz Steppes,[1] between the Urals and the Altai, in the Karkaralinsk and Akmolinsk districts. Very rich copper ores, consisting of malachite, azurite, and bornite, have been found here in sandstones reported to be of Paleozoic age. At Nankat, west of Kokand in Turkestan, deposits of metallic copper have been discovered in sandstones and gypsiferous marls of Tertiary age; fossil wood and chalcocite are also found.[2]

The lower Permian (Rothliegende) of Bohemia,[3] along the south slope of the Riesengebirge, contains similar ores.

Over a large part of western Europe the Triassic is copper-bearing, and together with the copper more or less lead is found.

In England, at Alderley Edge and Mottram St. Andrews,[4] south of Manchester, copper ores have been mined. They occur in the cement of Triassic sandstones and conglomerates and consist of copper carbonates, galena, pyromorphite, and vanadinite; also some barite, manganese, and cobalt. The ores are said to contain at most 1.4 per cent. copper. The mineral mottramite, a vanadate of copper and lead, was discovered at this place and vanadium was extracted from the ores.

In Germany the Triassic is divided into three parts—the lower Variegated Sandstone ("Buntsandstein"), the middle Shell Limestone (Muschelkalk); and the upper marls and sandstones (Keuper); of these the lower and upper divisions contain lead and copper ores.

In Bavaria the Keuper contains galena and chalcopyrite in certain gypsiferous beds, and these minerals are associated with a little zinc blende and barite.

In Württemberg galena with barite and some oxidized copper ores is generally distributed in the Corbula bed of the lower, gypsiferous Keuper. In the Palatinate, near Freihung, the littoral characteristics of the formation are plainly indicated and there is an abundance of fossil wood; at two horizons the sandstones contain galena and cerussite and were formerly worked.

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[1] A. Addiassewich, A journey to central Asia, *Trans.*, Inst. Min. and Met., vol. 17, 1907–1908, pp. 498–522.

[2] R. Beck, Lehre von den Erzlagerstätten, 1909, vol. 2, p. 172.

[3] F. Gürich, *Zeitschr. prakt. Geol.*, 1893, pp. 370–371.

[4] Phillips and Louis, Ore deposits, 1896, pp. 266–269.

In the "Buntsandstein" in Prussia and Lorraine, near Saarlouis and other places, a formation known as the Voltzia sandstone is particularly rich in lead and copper ores, which at times have been mined. The bed contains abundant plant remains. The minerals are cerussite, galena, chalcocite (?), and copper carbonates.

The best-known deposits of the European Triassic are those of Commern and Mechernich, not far from Aix-la-Chapelle, in Prussia. Lead ores have been mined here for several hundred years, but it is reported that the mines may soon be closed. The ores are of low grade and are mined in open cuts by removing about 130 feet of overburden. In 1903 the ores averaged 1.5 per cent. lead. The ore minerals are galena and cerussite, with a little chalcopyrite and barite, the latter filling veins and veinlets in the sandstone. Small amounts of silver, nickel, and cobalt are present. The thickness of the ore-bearing sandstone is about 20 meters. The general occurrence of the galena in so-called "Knoten" or knotty concretions is very remarkable. They often enclose several sand grains and some of them are bounded by the crystal faces of the galena. The epigenetic character of the ore is beyond doubt.

American Occurrences.[1]—On the North American continent copper ores are widely distributed in the "Red Beds" of the Southwest, in Texas, Oklahoma, New Mexico, Arizona, Colorado,

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[1] E. T. Dumble, *First Ann. Rept.*, Geol. Survey Texas, 1889, p. 186.

E. J. Schmitz, Copper ores in the Permian of Texas, *Trans.*, Am. Inst. Min. Eng., vol. 26, 1896, pp. 1051–1052.

S. F. Emmons, Copper in the Red Beds, *Bull.* 260, U. S. Geol. Survey, 1905, pp. 221–232.

W. H. Emmons, The Cashin mine, *Bull.* 285, U. S. Geol. Survey, 1906, pp. 125–128.

E. P. Jennings, *Trans.*, Am. Inst. Min. Eng., vol. 34, 1904, p. 839.

H. W. Turner, *Trans.*, Am. Inst.Min. Eng., vol. 33, 1903, p. 678.

W. Lindgren, L. C. Graton, and C. H. Gordon, The ore deposits of New Mexico, *Prof. Paper* 68, U. S. Geol. Survey, 1910.

H. S. Gale (Idaho), *Bull.* 430, U. S. Geol. Survey, 1909, pp. 112–121.

W. Lindgren (Colorado), *Bull.* 340, U. S. Geol. Survey, 1907, pp. 170–174.

W. A. Tarr (Oklahoma), *Econ. Geol.*, vol. 5, 1910, pp. 221–226.

A. E. Fath (Oklahoma), *Econ. Geol.*, vol. 10, 1915, pp. 140–150.

L. M. Richard (Texas), *Econ. Geol.*, vol. 10, 1915, pp. 634–650.

A. F. Rogers, Origin of copper ores of the "Red Bed" type, *Econ. Geol.*, vol. 11, 1916, pp. 366–380.

Wyoming, Utah, and Idaho and, though conspicuous by green and blue colors, rarely prove of economic importance.

The ore occurs in arkose sandstone, conglomerate, or clay shale and is usually associated with plant remains and fossil wood. These strata were accumulated in shallow seas or as subaerial deposits by a process of rapid degradation of adjacent land areas of the Rocky Mountain region, and they have been referred to the upper Carboniferous, Permian, Triassic, and Jurassic.

In Texas the copper-bearing beds appear over large areas in Permian sandstones and shales. They lie at several horizons, in strata rich in plant remains; covellite, chalcocite, chalcopyrite, and pyrite are the minerals mentioned. The area is said to extend from 33° to 34° in latitude and from 98° to 100° in longitude.

Fig. 128.—Chalcocite nodules replacing fossil wood and carbonaceous shale of "Red Beds," Red Gulch, Colorado. Natural size.

In Oklahoma nodules of chalcocite are found in red shales and sandstones of the same age. Fossil wood is often converted to chalcocite, sometimes with a shell of chalcopyrite; unusually high silver assays of 31 ounces per ton and traces of gold are reported.

In Colorado these ores have been recorded at several places, notably at Red Gulch, Fremont County, where Lindgren observed nodules of chalcocite with barite in black carbonaceous shale; sections (Fig. 128) show that the copper sulphide actually replaces the coal and shale. The horizon is probably the uppermost Carboniferous. In northeastern Arizona, according to Gregory,[1] small quantities of oxidized copper ores are frequently seen

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[1] H. E. Gregory, *Prof. Paper* 93, U. S. Geol. Survey, 1917, p 140
J. M. Hill, *Bull.* 540, U. S. Geol. Survey, 1913, p. 163.

in the La Plata sandstone. North of the Colorado River they appear in the Carboniferous of the Kaibab Plateau. S. F. Emmons believed that the copper in the oxidized ore and chalcocite in the Aubrey limestone near Grandview, Arizona, was leached from the "Red Beds" and carried down into the limestone.

In southwestern Colorado copper, often accompanied by vanadium ores, is widely distributed in the Jurassic La Plata sandstone. W. H. Emmons has described the Cashin vein in this formation near Placerville. The ores are here argentiferous chalcocite, covellite, and bornite, with some calcite. No igneous rocks are present and Emmons believes that the ores were leached from the "Red Beds." There is an active circulation of water in the formation, and springs with salt, sulphates, and hydrogen sulphide abound. A production of about 300,000 ounces of silver and 700,000 pounds of copper is recorded from this mine.

The greatest development of the copper-bearing sandstones is in New Mexico; considerable production from picked ore has been achieved at the Nacimiento deposits, in the northern part of the State, where the "Red Beds" rest on pre-Cambrian granitic rocks which contain much older copper deposits. The beds have been referred to the Triassic on the evidence of fossil plants. According to Schrader[1] most of the copper ores occur in the basal beds and are confined within a thickness of 25 feet in a reddish-white sandstone rich in fossil wood, which is largely chalcocitized. A tree trunk 60 feet long with a basal diameter of 2½ feet is mentioned, which was almost wholly converted to copper glance. Besides malachite, azurite, and chrysocolla, there is some barite and, at one place, cerussite. The low-grade ores have not been utilized.

According to the same geologist the copper-bearing beds of the Zuni Mountains, in northeastern New Mexico,[2] lie at the base of the "Red Beds," resting on pre-Cambrian gneisses which contain older copper veins. The sandstones, shales, and marls for 30 to 60 feet just above the base of the beds contain oxidized ores and chalcocite replacing wood.

Graton describes in detail the ores from the Tecolote district, San Miguel County, which are partly in the "Red Beds" of the upper Carboniferous (Abo formation), and partly at a higher

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[1] F. C. Schrader, *Prof. Paper* 68, U. S. Geol. Survey, pp. 141–149.

[2] *Idem*, p. 134.

horizon, perhaps in the Dakota sandstone. The calcareous cement of the arkose is replaced by chalcocite, bornite, chalcopyrite, and pyrite, the replacement extending into the feldspar grains.

In the Oscura Range, also in New Mexico, red sandstones, probably Carboniferous "Red Beds," contain chalcocite, bornite, and chalcopyrite, in part as replacement of fossil wood. Turner mentions the occurrence of plant remains, said to have been identified as the Triassic *Podozamites crassifolia*, the same cycad which is characteristic of the deposits at Abiquiu first studied by Newberry.

Graton believes, contrary to Turner, that the copper ores have been introduced into the strata by waters ascending along a number of dislocations in the sandstone.

Finally, H. S. Gale describes copper ores from southern Idaho which occur in the Ankareh maroon shales and sandstones of the Triassic or Carboniferous (equivalent to the Permo-Carboniferous of the Fortieth Parallel Survey). (See Fig. 10.) A thick limestone (including the Meekoceras beds) underlying these shales is believed by some geologists to be Triassic.

There are then at least two main cupriferous formations in the Southwest—(1) the upper Carboniferous "Red Beds," equivalent to the Permo-Carboniferous, or the Abo formation; (2) the undoubtedly Jurassic La Plata sandstone.

The silver deposits in the supposedly Triassic sandstones of Silver Reef,[1] in southern Utah (Harrisburg district), which created a boom about 1880, are now worked only on a small scale. The ores were silver chloride above the water level and native silver and argentite in depth; copper was also present, and selenium is reported. Plant remains were abundant. A secondary concentration from a primary argentiferous chalcocite is the probable genesis.

In Nova Scotia, Cumberland County, chalcocite nodules, with remains of pyrite, and also chalcocitized wood, are found in the Permian sandstone.

[1] C. M. Rolker, The silver sandstone district of Utah, *Trans.* Am. Inst. Min. Eng., vol. 9, 1881, pp. 21–33.

J. P. Rothwell, The silver sandstone formation of Silver Reef, *Eng. and Min. Jour.*, vol. 29, 1880, pp. 25, 48, 79.

J. S. Newberry, Report on the property of the Stormont Silver Mining Company, *Eng. and Min. Jour.*, vol. 30, 1880, p. 269; vol. 31, 1881, pp. 4–5.

J. F. Kemp, Ore deposits of the United States, 1900, p. 334.

South America.—The well-known and long-worked copper deposits of Coro-Coro,[1] in Bolivia, are contained in a series of sandstones, believed to be of Permian age. There are several cupriferous beds with disseminated native copper, in places dendritic, and much gypsum, also some native silver, domeykite, and chalcocite. The copper-bearing beds are much lighter in color than the prevailing deep-red sandstones.

According to Steinmann the strata are of Cretaceous age and the copper was introduced by hot waters derived from an intrusion of diorite. Nevertheless the descriptions suggest strongly that the deposits belong in a different class.

Africa.—Sufficient information is not at hand to decide whether the recently opened Katanga ores[2] of southeastern Belgian Kongo, near Rhodesia, belong to the class of deposits described in this chapter. Large masses of high-grade oxidized copper ores are contained in sandstones, shale, and limestone, probably of Paleozoic age. The ores are of high grade (8 to 12 per cent. copper) and are said to contain a little gold and silver; some manganese, cobalt, and nickel are present. Barite and quartz appear as gangue minerals.

Genesis of Sedimentary Copper Ores.—The epigenetic character of the copper deposits in sandstone is proved beyond reasonable doubt. The replacement of coal, carbonaceous shale, and calcareous or kaolinic sandstone cement by chalcocite is also proved. The gangue minerals are few and quartz is conspicuously absent. Barite in small amounts is rather common. Irregularity in dissemination is typical, though the ores often follow certain horizons rather persistently. The entire independence of the occurrence of igneous rocks is marked.

The occurrences are mainly on the flanks of older continental areas containing pre-Cambrian copper deposits; the sandstones were rapidly laid down as arkoses, indicating a long epoch of rock decay, the products of which were swept away during a following arid epoch. Considering the evidence as a whole the sedimen-

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[1] Older literature: See Stelzner and Bergeat, Die Erzlagerstätten, vol. 1, 1904, p. 419.

G. Steinmann, Rosenbusch Festschrift, 1906, pp. 335–368.

F. C. Lincoln, *Min. and Sci. Press*, Sept. 29, 1917.

Lester W. Strauss, *Min. Mag.*, vol. 7, 1912, p. 207.

[2] S. H. Ball and M. K. Shaler, Mining conditions in the Belgian Congo, *Trans.*, Am. Inst. Min. Eng., vol. 41, 1911, pp. 189–219; also *Econ. Geol.*, vol. 9, 1914, pp. 617–632, with literature.

tary deposits must have contained finely divided copper ores, in part from solutions derived from the land area, in part as cupriferous detritus. When atmospheric waters charged with salt and gypsum searched these beds they must have taken this copper into solution and concentrated it at certain horizons when reducing substances like coaly vegetable matter were available. In most cases the solution probably contained the copper as sulphate, though where much salt was present it might well have been transformed into chloride.

Genesis by ascending thermal solutions of meteoric origin is a possible cause of some deposits.

Graton, Fath and Rogers have noted pyrite, bornite and chalcopyrite in the chalcocite nodules and the latter two authors have shown that the chalcocite replaces earlier pyrite or marcasite (Fig. 128). No doubt this is true in places but this view may not be universally applicable. Rogers[1] believes that the fossil wood was successively replaced by hematite, pyrite, bornite and chalcocite, a rather improbable series of events, considering that the wood structure is preserved even in the fourth and last replacement. Graton finds sharp cubes of pyrite in the chalcocite and no evidence of replacement of pyrite.

In the precipitation the most important chemical reactions were those between the hydrocarbons of plant remains and the calcareous cement or the kaolin in the sandstone on one hand and the cupriferous solutions on the other hand. How the metallic copper in these ores was precipitated is not known. The Bolivian occurrences show distinct bleaching of the reddish sandstone around the copper aggregates, from which it may be inferred that the solutions were reducing in character.

VANADIUM AND URANIUM ORES IN SANDSTONES[2]

Composition.—Many of the copper deposits described above carry some vanadium as vanadinite or volborthite. Lately vanadium with some uranium and a trace of radium has been

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[1] *Op. cit.*

[2] G. P. Merrill, Non-metallic minerals, 1904, pp. 299–320.

W. F. Hillebrand and F. L. Ransome, Carnotite, etc., in western Colorado, *Am. Jour. Sci.*, 4th ser., vol. 10, 1900, pp. 120–144. *Bull.* 262, U. S. Geol. Survey, 1905, pp. 9–13.

H. Fleck and W. G. Haldane, *Rept.* State Bureau of Mines, Colorado, 1907, pp. 47–115.

shown to be common in certain Jurassic sandstones in Colorado and Utah. The deposits in western Colorado are now worked and a reduction plant is located at Vanadium, near Placerville.

A number of unusual minerals are contained in these deposits. One of the most conspicuous is *carnotite* ($2U_2O_3.V_2O_5.K_2O.3H_2O$) a crystalline potassium-uranium vanadate, first named by Fuchs and Cumenge, which forms a bright yellow powder occurring in seams and on fossil wood. An analysis by W. F. Hillebrand gave:

	Per cent.		Per cent.
UO_3	54.89	CuO	0.15
V_2O_5	18.49	MoO_3	0.18
CaO	3.34	H_2O	4.54
BaO	0.90	CO_2	0.56
K_2O	6.52	Insoluble	7.10
PbO	0.13		

In calcio-carnotite potassium is replaced by calcium.

Associated with carnotite are a number of other obscure vanadium minerals, which appear as earthy black, brown and red coatings or fissure fillings. They are crystalline and highly hydrous vanadates. *Metahewettite*[1] ($CaO.3V_2O_5.9H_2O$) is a dark red calcium vanadate containing, in per cent., $70V_2O_5$, 7.25CaO and $21.30H_2O$.

Pintadoite is another mineral of similar composition but with only 42.4 per cent. V_2O_5; it forms green efflorescences and occurs in Utah.[2] *Uvanite*, a brownish-yellow uranium vanadate ($2UO_3$.-

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J. M. Boutwell, *Bull.* 260, U. S. Geol. Survey, 1905, p. 205.

H. S. Gale (Carnotite in Colorado), *Bull.* 340, U. S. Geol. Survey, 1908; *Idem, Bull.* 315, 1906, pp. 110–117.

F. L. Hess, Vanadium deposits in Colorado, Utah, and New Mexico, *Bull.* 530, U. S. Geol. Survey, 1912.

Idem, Mineral Resources, Annual issue, 1912, pp. 1003–1036.

F. L. Hess, A hypothesis for the origin of the carnotites, *Econ. Geol.*, vol. 9, 1914, pp. 675–688.

R. B. Moore and K. L. Kitthil, A preliminary report on uranium, radium and vanadium, *Bull.* 70, U. S. Bureau of Mines, 1913.

K. L. Kitthil and John A. Davis, Mining and concentration of carnotite ore, *Bull.* 103, U. S. Bureau of Mines, 1917.

Parsons, Moore, Lind and Schaefer, Extraction and recovery of radium, uranium and vanadium, etc., *Bull.* 104, U. S. Bureau of Mines, 1915.

[1] W. F. Hillebrand, H. E. Merwin and F. E. Wright, *Proc.*, Am. Philos. Soc., vol. 53, 1914, pp. 31–54.

[2] F. L. Hess and W. T. Schaller, *Jour.*, Washington Acad. Sci., vol. 4, pp. 576–579.

$3V_2O_5.15H_2O$) with 39.60 per cent. UO_3 and 37.70 per cent. V_2O_5 is mined in Emery County, Utah.

Volborthite and *calcio-volborthite* both vanadates of copper, and a uranium sulphate have also been identified.

Roscoelite, a dark green vanadium mica is abundant as veins and as replacements in the cement of some sandstones (Fig. 129) at Placerville, Colorado. About two-thirds of the aluminum is replaced by vanadium so that it contains from 20 to 29 per cent. V_2O_3. Some sandstones contain as much as 20 per cent. of roscoelite.[1]

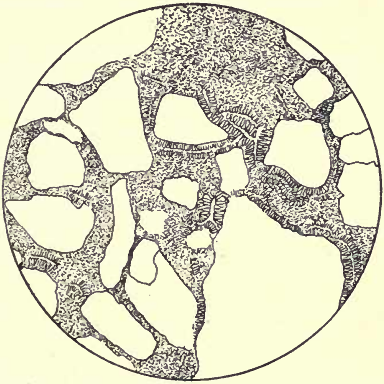

FIG. 129.—Vanadium ores in sandstone. White areas, quartz; shaded areas, roscoelite, partly radial. Magnified 25 diameters. *After F. L. Hess, U. S. Geol. Survey.*

Chromium is also present in these remarkable ores, probably as *mariposite,* or chromium mica. Barium, copper, lead, molybdenum and arsenic are contained in the ores as shown by the analysis of carnotite. Molybdenum is quite abundant in some places and appears to occur as a soluble sulphate—the dark blue ilsemannite. A similar occurrence of the latter mineral is reported from South Africa. Native selenium has been identified by Hillebrand, associated with metahewettite. Some of these minerals are quite certainly secondary products.

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[1] Roscoelite is also known from some gold-quartz veins and is often intimately associated with native gold. Mariposite is a characteristic mineral of many gold quartz veins.

The carnotite, which is the most abundant uranium mineral, contains a small trace of radium[1] which is recovered. Gypsum is about the only gangue mineral associated with the ores.

The ores are not rich. The carnotite ores contain about 1.5 to 3 per cent. U_5O_8 and 3 to 5 per cent. V_2O_3. Concentration has been attempted in some cases.

The roscoelite ore at Placerville contains about 3.50 per cent. V_2O_3 and 0.05 per cent. UO_3.

Occurrence.—The ores are found in the Plateau province of horizontal or gently inclined strata in southwestern Colorado and eastern Utah. The best known localities are at Placerville, Colorado, and in the LaSal, Paradox and Sindbad valleys somewhat farther west.

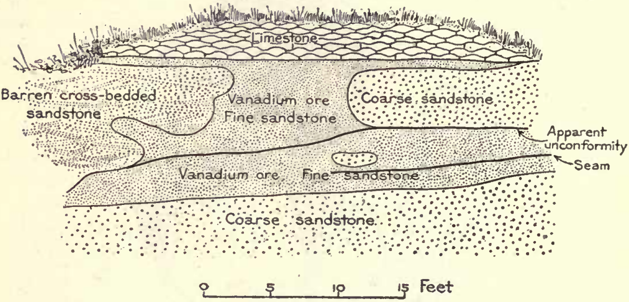

Fig. 130.—Sketch of vanadium-bearing sandstone at mine of Primos Chemical Company, on the east side of Bear Creek, Newmire, Colo. *After F. L. Hess, U. S. Geol. Survey.*

The ores are mainly confined to the McElmo and LaPlata formations of white, often cross bedded Jurassic sandstone which frequently contain much transported partly carbonized wood. They follow certain horizons or appear in fissures of flat veins or in brecciated zones (Fig. 130) and are often associated with the fossil wood.

Some observers have thought the ores merely superficial but it now seems certain that they may be found in depth. At Placerville the workings are said to have penetrated 2,000 feet under ground in horizontal direction.

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[1] At the rate of one gram of radium in 3,000 kilograms of metallic uranium.

Not similar to those deposits, yet perhaps of a similar origin, are the recently discovered important vanadium deposits at Minasragra,[1] near Quisque, Province of Pasco, Peru, described by D. F. Hewett and W. F. Hillebrand. The vanadium sulphide, patronite, occurs here on a large scale as greenish-black masses associated with a hydrocarbon and a peculiar nickel-bearing pyrite. An analysis of the patronite gave 58.79 per cent. sulphur, 19.53 per cent. vanadium, 0.18 per cent. molybdenum, 1.87 per cent. nickel, and 3.47 per cent. carbon. The deposit is said to be a vein with much bitumen and clay in gently dipping Cretaceous strata. Igneous rocks are present in abundance.

Genesis.—Traces of vanadium are found, according to Hillebrand, in most igneous rocks, and some varieties of augites carry as much as 2 per cent. of the metal. Titanic iron ores usually contain a fraction of a percent. Smaller quantities occur as oxidation products in many ore deposits, mostly as vanadinite or descloizite. Vanadium tends to concentrate in clays and shales; it is also concentrated in coal; the ashes of many varieties are rich in vanadium.[2] A coal from the copper deposits in sandstone of Red Gulch, Colorado, contained, according to Hillebrand,[3] 0.18 per cent. vanadium. The processes of weathering and vegetation appear to be favorable to its concentration; and, to some extent, uranium shares in this behavior.

F. L. Hess believes that adjoining land areas may have contained uranium- and vanadium-bearing veins at a certain level which would account for the ores being practically confined to one stratigraphic horizon, but such an assumption is scarcely supported by any facts. At present the localization of the ores cannot be satisfactorily explained. The deposits are probably products of concentration, by meteoric waters, of small quantities of the metals distributed in littoral beds or in land deposits and derived from older deposits of some kind in ancient land areas of igneous or metamorphic rocks.

Production and Use.—In 1915, 47,000 tons of ore were mined in Colorado containing, according to U. S. Geological Survey,

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[1] D. F. Hewett, Vanadium deposits of Peru, *Trans.*, Am. Inst. Min. Eng., vol. 40, 1909, pp. 274–299.

W. F. Hillebrand, The vanadium sulphide, patronite, *Jour.*, Am. Chem. Soc., vol. 29, 1907.

W. F. Hillebrand, *Am. Jour. Sci.*, 4th ser., vol. 24, 1907, p. 141.

[2] F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 705.

[3] *Bull.* 340, U. S. Geol. Survey, 1908, p. 172.

19.9 tons of metallic uranium, 627 tons of vanadium (mainly from roscoelite ores) and 6.1 grams of radium, the value of the metals being about $700,000. Some years ago most of the uranium ores were exported and radium extracted abroad. In 1914, the ores mined contained 87.2 tons of uranium and 22.3 grams of radium. The pure vanadium ores of Placerville are roasted with sodium chloride, the resulting sodium vanadate extracted with water and precipitated with ferrous sulphate as iron vanadate which is shipped east for reduction to ferro-vanadium. About 1 per cent. vanadium added to steel increases its toughness and resistance to torsion and high temperature. It is, however, less essential to the steel industry than tungsten, and the principal supply is obtained from the Peruvian patronite mine. The value of ferro-vanadium alloy is about $1,000 per ton. Minor amounts of vanadium salts are used as mordants for dyeing and cloth printing and for other chemical purposes. Various ore deposits yield small quantities of vanadium ores such as vanadinite and descloizite. In the carnotite ores there is difficulty in separating uranium from vanadium and only a small price is paid for the latter.

Uranium salts have a limited use for a yellowish-green glass and for pottery; also as a mordant in dyeing. Ferro-uranium is at present not used in the steel industry. Radium is separated from uranium by a complicated process explained in *Bull.* 104, U. S. Bureau of Mines. It is produced as a chloride or bromide and its principal use is in medical science, various diseases yielding more or less to its emanations. It is said that it can be extracted at a cost of $37 per milligram.

Radium in corresponding quantities is also contained in uraninite (crystalline) and pitchblende (amorphous), both essentially $UO_2.UO_3$ with 80 ± per cent. uranium oxides.[1] It is obtained in small quantities from gold-pyrite veins of Gilpin County, California, from cobalt arsenide veins of Joachimsthal, Bohemia, and from tin veins of Cornwall, all of which are deposits formed at higher temperatures. Various uranium minerals also occur in pegmatite dikes.

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[1] E. S. Bastin and J. M. Hill, *Prof. Paper* 94, U. S. Geol. Survey, 1917, pp. 121–128.

THE COPPER-BEARING SHALES OF MANSFELD[1]

It seems proper to consider at this place the celebrated cupriferous shale (Kupferschiefer) of Mansfeld, in central Germany, for, though not identical with the deposits described in this chapter, it presents most interesting analogies to them.

A flourishing mining industry is still based on the Kupferschiefer, the annual production being approximately 700,000 metric tons of ore containing between 2 and 3 per cent. of copper.

The first stratum deposited in the subsiding basin of the upper Permian in central Germany was a marine conglomerate of slight thickness. "Above it extends like a black shroud the thin bed of cupriferous shale, one of the most remarkable products of the geologic ages. Characterized by its fauna as a shallow-sea deposit, full of plant remains carried in from adjacent coasts, the formation bears the stamp of an organic mud deposit admixed with organic precipitates.[2]

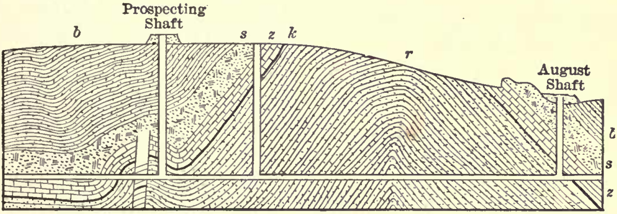

FIG. 131.—Section of a part of the copper-bearing shale bed at Mansfeld, Germany. *After Schrader.*

Above the cupriferous shale, which is less than 1 meter thick, extends a marine limestone (Zechstein), 6 to 10 meters thick, and above that lie the great gypsum and salt masses of the uppermost Permian. Folding and faulting have since affected the beds, and the mining now follows the inclined strata to a depth of 500 meters. The cupriferous bed averages 50 centimeters in thickness, but only the lower 20 to 30 centimeters is utilized as ore (Fig. 131).

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[1] Best description with literature in Stelzner and Bergeat, Die Erzlagerstätten, vol. 1, 1904, pp. 391–417.

[2] F. Beyschlag, in *Deutschlands Kalibergbau,* 1907, p. 4.

The ores are sulphides, in minute distribution through the shale, giving it a bronzy appearance. Chalcopyrite predominates, but there are also bornite, pyrite, chalcocite, and rarely galena and tetrahedrite. Small quantities of nickel, cobalt, selenium, vanadium, and molybdenum have been recognized; there is also about 0.015 per cent. of silver. Zinc is present, and in the upper part of the bed, not mined, there is more zinc than copper. An average analysis is as follows:[1]

AVERAGE ANALYSIS OF CUPRIFEROUS SHALE

Dr. Haase, Analyst

	Per cent.		Per cent.
SiO_2	33.15	Ag	0.014
Al_2O_3	17.3	Ni	0.018
CaO	10.4	S	2.31
MgO	1.0	CO_2	9.24
Fe	2.6	H_2O	1.7
Zn	1.276	Bitumen	9.06
Cu	2.75		

There is about 3 per cent. K_2O and 1 per cent. Na_2O; lead amounts to 1.50 per cent., manganese to about 0.25 per cent.

There is practically no gangue, except veinlets of gypsum and barite. The bed is cut by faults, along which in places occur barite, anhydrite, calcite, niccolite, pyrite, and chalcopyrite, and near these breaks (called "Rücken") the metal content is subject to enrichment, impoverishment, or removal upward in adjacent beds. Bergeat declares that these changes take place on secondary fissures and cracks.

There has been much controversy about the Mansfeld deposits. The majority of geologists regard them as sedimentary and syngenetic: von Groddeck, Stelzner, Freiesleben, and von Cotta held this view, and it is shared by Bergeat. Posepny and Beck believe them epigenetic and think that the metals were probably introduced in the shale from the Rücken.

The Kupferschiefer is certainly not an ordinary marine deposit precipitated from the sea water.[2] It was laid down in a shallow sea which was full of decaying vegetable and animal remains and into which were probably discharged cupriferous

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[1] Stelzner and Bergeat, Die Erzlagerstätten, vol. 1, 1904, p. 396.

[2] Sea water contains a trace of copper, as shown by Dieulafait (*Ann. chim. phys.*, 5th ser., vol. 18, 1879, p. 359; *C. R.*, 90, p. 1573; 96, p. 70; 101, p. 1297) and others, but the amount present seems utterly insufficient to account for the Mansfeld deposits.

waters from the surrounding littoral, most likely sulphate solutions derived from the eruptives and the ore deposits of the early Permian epochs. No one can read the description of the great uniformity of distribution without being impressed with the very strong arguments for a syngenetic origin.

The characteristic presence of nickel, cobalt, vanadium, and selenium recalls the epigenetic deposits in sandstone so abundant around the shores of the Permian sea, in Bohemia and Russia, for instance. The Mansfeld basin was simply, then, the final collecting place of the solutions derived from adjacent desert shores.

COPPER SULPHIDE VEINS IN BASIC LAVAS

General Features.—All basic lavas contain copper, but in many cases conditions were evidently unfavorable for the concentration of copper immediately after the eruption, and the rocks retained their copper until later opportunities for ore formation were offered. The existence of vast masses of such basic lavas near the surface, without any indication of copper concentration (*e.g.*, the Columbia River lava or the basalts of the Hawaiian volcanoes), shows plainly that the ordinary surface waters at slight depth are not competent to dissolve and concentrate accessory metals contained in these rocks. A depth of perhaps a few thousand feet seems to be necessary, under the most favorable conditions, for waters of meteoric origin to extract the copper; though it is, of course, possible that such waters, when ascending in suitable channels, may deposit the dissolved copper at higher horizons. In some of the veins here discussed epidote is present, but more frequently it is absent, and the veins then assume the well-known type of chalcopyrite in a quartz-calcite-siderite gangue. Such veins, deposited by ascending surface waters of the deeper circulation, may not be easy to distinguish from those whose development is a phase of intrusive after-effects. Nor can it be denied that in these veins may be concentrated some gold and silver from the igneous rock; in general, however, they will be found much poorer in gold and silver than deposits connected with the intrusive processes.

Whether native copper, bornite, or chalcopyrite will form seems to be dependent upon the quantity of sulphur which the lavas were able to retain at their eruption.

The Nikolai Greenstone.—The copper deposits in the Nikolai greenstones of the Copper River region, described by F. C. Schrader, W. C. Mendenhall, A. C. Spencer, and lately again by F. H. Moffit,[1] are of special interest. Flows of Triassic or Carboniferous basalts about 4,000 feet in thickness are covered by 2,000 feet of Triassic limestone, which in turn is overlain by a thick series of Jurassic strata. The latter are cut by monzonitic intrusives, which are accompanied by a different kind of metallization characterized by prominent gold deposits.

The Nikolai greenstones are amygdaloid flows of basalt; the amygdules contain scarcely any zeolites, but are filled with chlorite, chalcedony, and quartz and carry no copper. Copper sulphides are extremely common in the flows, but occur in slips, brecciated zones, and faults. The minerals are chalcopyrite, pyrite, and bornite, with calcite and a little quartz; there is some epidote, not always present.

One of the fissure zones extends up into the Triassic limestone above the greenstone. In the latter a little bornite and chalcocite appears and the zone cuts across an older series of quartz-epidote veins carrying the same two sulphides with a little native copper. In the limestone the fissure zone develops into the remarkable and valuable deposit worked in the Bonanza mine. It is an almost solid body of massive chalcocite with conchoidal fracture, traced for 400 feet and with a greatest width of 25 feet; its depth is apparently limited. There are no gangue minerals and the limestone adjoining the chalcocite is not altered. No intrusive rocks are present.

It is probable that the ores characteristic of the Nikolai greenstones are derived from the rock itself. The copper deposits seemed to be formed mainly after the Triassic limestone had been laid down, and it is likely that meteoric waters did the work. The waters must have descended through limestones and shales in which they would have acquired chlorides, sulphates, carbonates, carbon dioxide, and hydrogen sulphide, and they would therefore be competent to dissolve copper from the greenstones which they traversed. The chalcocite alters superficially to covellite and copper carbonates; there is no evidence that it has replaced pyrite and it may have been deposited in its present form.

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[1] F. H. Moffit and S. R. Capps, *Bull.* 448, U. S. Geol. Survey, 1911.
F. H. Moffit, *Bull.* 662, U. S. Geol. Survey. 1917. pp. 155–182.

C. F. Tolman[1] has recently found remnants and structures indicating bornite and believes that the chalcocite is secondary and mainly derived from bornite. The final word about this interesting deposit has not been said.

COPPER SULPHIDE VEINS IN INTRUSIVE BASIC ROCKS

Veins containing pyrite and chalcopyrite, occasionally with other sulphides, in a gangue of quartz, calcite, dolomite, and siderite, more commonly quartz alone, occur abundantly in intrusive basic rocks, such as diabase or gabbro. Here, however, the distinction between the results effected by water of atmospheric origin and those effected by magmatic solutions becomes increasingly difficult.

F. E. Wright[2] has pointed out the fact that the intrusive Keweenawan gabbro of Mount Bohemia contains veins with chalcopyrite, bornite, chalcocite, calcite, and quartz, while in the surface lavas of the same series native copper is the principal ore mineral. This seems an excellent illustration of the retention of volatile sulphur by intrusives, contrasted to its escape from the extrusive flows. The origin of the water which was the vehicle of deposition in these veins may remain an open question.

Along the foot-hills of the Sierra Nevada of California extends a belt of andesitic rocks of Jurassic age collectively called "greenstones." They consist of massive and schistose rocks, including surface flows, tuffs, and intrusions mixed. Within this belt, for instance in Yuba and Nevada counties, short and irregular quartz veins with pyrite and chalcopyrite are common. Probably these veins derived their copper from the greenstones, and undoubtedly they were formed at a time when the rocks now exposed were covered by several thousand feet of overlying and now eroded igneous rocks.

Other deposits, such as those at the Dairy Farm in Placer County and at Campo Seco in Calaveras County, are, according to A. Knopf,[3] replacement deposits along shear zones in the same belt of amphibolites and other greenstones. The minerals are pyrite and chalcopyrite, with a trifle of galena and zinc blende,

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[1] *Trans.*, Am. Inst. Min. Eng., vol. 54, 1917, pp. 402–441.

[2] F. E. Wright, The intrusive rocks of Mount Bohemia, Michigan, *Seventh Ann. Rept.*, Geol. Survey Michigan, 1908.

[3] A. Knopf, Notes on the foot-hill copper belt of the Sierra Nevada, *Bull.* 17, California Univ. Dept. Geology, vol. 4, 1906, pp. 411–421.

associated with quartz, calcite, epidote, chalcedony, and sometimes zeolites. The sulphides contain a little silver and a trace of gold. Similar are the so-called "iron belts" of pyrite and chalcopyrite in the Ophir mining district of gold-silver quartz veins in Placer County.[1] They are contained in amphibolite, but do not always extend parallel to the schistosity; the width varies from a few feet to 400 feet, the length is in places half a mile. No shear planes are visible along these zones. They enrich the distinctly later gold-quartz veins which cross them. The sulphides are associated with the amphibole, epidote, feldspar, and quartz of the amphibolites and are often intergrown with magnetite or ilmenite. In the paper cited they were interpreted as concentrations of copper from the surrounding rocks, formed by chemical action during the progress of the dynamic metamorphism which produced amphibolites from primary diabasic rocks.

Other and much larger copper deposits are found in the same region, on the north in Shasta County and on the south in Calaveras County, but at both places the evidence points clearly to an origin by solutions derived directly from the magma.

In the Encampment district, Wyoming, A. C. Spencer[2] studied deposits of primary chalcopyrite enriched by chalcocitization, and probably of pre-Cambrian age. The metallization is localized in shattered zones in quartzite, or between quartzite and schist, close to intrusive gabbro or diorite which contains copper (p. 9), sometimes visible as chalcopyrite.

Spencer gives several good structural reasons indicating that the depositing waters were ascending and believes that the copper was leached from the cupriferous gabbro. The minerals present do not indicate especially high temperatures. Although the deposits were formed at considerable depth, as shown by the flexing of the schist bands, the quartzite was decidedly in its zone of fracture.

OTHER VEINS DEPOSITED BY WATERS OF THE UPPER CIRCULATION

In the preceding pages it has often been pointed out that the competency of the circulation of certain kinds of atmospheric

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[1] W. Lindgren, *Fourteenth Ann. Rept.*, U. S. Geol. Survey, 1895, pp. 262–264.

[2] The copper deposits of the Encampment district, Wyoming, *Prof. Paper* 25, U. S. Geol. Survey, 1904.

waters to form many mineral deposits cannot be questioned and that it may be difficult or impossible to determine the origin of certain occurrences.

Nevertheless, the fact stands firm that surface waters of the ordinary type, even in slightly heated ascending currents, do not form mineral deposits even in localities where the conditions are such that they might be expected to do so, as in the Alps, for instance. There are, however, other localities, particularly in the region of the saline Paleozoic and Mesozoic beds of central Germany, where such deposition appears to have taken place. Veins of this origin are likely to contain an abundant gangue of calcite, dolomite, or barite; with some quartz and a scant amount of sulphides.[1]

The sweeping generalizations of F. Hornung and his interpretation of all the mineral veins of the Harz Mountains as being formed by inter-Permian brines cannot be accepted, but it is not improbable that he is correct in believing that many barite and hematite veins have had this origin.[2] In connection with this K. Ochsenius[3] showed that solutions containing 2.59 per cent. NaCl, 3.16 per cent. $MgCl_2$, and 1.85 per cent. $MgSO_4$ decompose chalcopyrite and chalcocite at room temperature. This action is slow and is noticeable only after several years. Galena was not dissolved.

Similar examples of ore deposition by saline waters also exist in the western part of the United States; one, the Cashin mine of Colorado, is mentioned above (p. 404). The prevailing influence of igneous intrusions on ore deposition is, however, so strong that it is difficult to establish the proofs of the less conspicuous deposition by purely meteoric water.

That the ordinary surface waters are in most cases quite incompetent to effect concentration is plainly shown by the lack of important mineralization in fissures and joints cutting the rocks of mining districts. In the Globe district, Arizona, for instance, the Paleozoic rocks are intersected by a network of dislocations which would offer excellent paths for these waters; and yet the important deposits are in no way connected with these fractures.

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[1] P. Krusch, Ueber die Zusammensetzung der Westfälischen Spaltenwässer, *Zeitschr. prakt. Geol.*, vol. 12, 1904, p. 252.

[2] F. Hornung, Ursprung und Alter des Schwerspates und der Erze im Harze, *Zeitschr.* Deutsch. geol. Gesell., vol. 57, 1905, pp. 291–360.

[3] *Idem*, p. 567.

Similarly "cross courses" often fault the gold-quartz veins of California and yet they are, as a rule, absolutely barren, often open fissures. Similar post-mineral fissures traverse lead and zinc veins in the Coeur d' Alene district, Idaho, but generally show no trace of mineralization.

The Cordilleran region contains many great ranges of pre-Cambrian rocks capped in places by Paleozoic and Mesozoic strata. Among them may be mentioned the Front Ranges of the Rocky Mountains in Colorado, the Wind River Range in Wyoming, and the Mission Range in Montana. Uplift, folding, and faulting have in each of these ranges intensified the circulation of meteoric waters, but in spite of this the ranges are remarkably poor in mineral deposits, which appear only in the vicinity of later intrusives. These relations show very plainly the slight concentrating power of ordinary cool surface waters and even of the waters of atmospheric origin that have become a part of the deeper circulation.

CHAPTER XXI

DEPOSITS RESULTING FROM REGIONAL METAMORPHISM

Rocks subjected to stress at moderate depths within the zone of fracture may rupture in closely spaced breaks, producing the appearance of a schistose structure. In such rocks no great chemical changes would occur, except perhaps by subsequent deposition along the tight fissures. At greater depth deformation may take place by granulation and recrystallization, accompanied by chemical changes that are effected by the aid of the scant rock moisture. Uralite and chlorite may form from pyroxene, the soda-lime feldspars may recrystallize to zoisite and albite, the quartz crystals may be crushed and elongated, new mica, particularly muscovite, may develop; also crystals of aluminum garnet. The chemical composition of the rock will, however, change but little; although the various transformations involve transportation of substance, this movement is not free, but is limited and hindered in all directions.

Under these circumstances it is improbable that processes of concentration could have much opportunity to assert themselves; the minute quantities of useful metals contained in the original rock could not easily assemble to form larger masses.

In amphibolite schist small grains of chalcopyrite, often intergrown with pyrrhotite and magnetite, appear to be more common than in the primary igneous rock from which the schist was derived. If even the slightest and slowest circulation of water was established during the deformation, some concentration of chalcopyrite could well take place, as it does in fissures traversing similar rocks.

When the deformation takes place at higher temperatures a number of minerals are developed which are similar or identical with those of contact metamorphism. It is often difficult, indeed, to draw the line between regional and igneous metamorphism, especially in intensely metamorphic regions where intrusive masses are abundant. There is reason to believe that at temperatures of, say, several hundred degrees some of the rocks, particularly limestones, become permeable to the gaseous emanations of water and metallic compounds yielded by intrusive masses, and thus an opportunity is afforded for the introduction of new substances which in places may become concentrated into ore deposits. To such a permeation in the deep zone of anamorphism many of the most enigmatic ore deposits of the crystalline schists may owe their origin. These deposits would then differ in some respects from the ordinary contact-metamorphic ores, which have, as a rule, developed only close to intrusive contacts, in most cases also actually within the zone of fracture.

Dissemination of sulphides is a phenomenon often encountered in almost any area of crystalline schists. In the majority of occurrences pyrite, pyrrhotite, and chalcopyrite are prominent; the sulphides of lead and zinc are far less common. Such disseminations are also particularly connected with amphibolitic or chloritic rocks. As indicated above, these ore minerals may have various modes of origin. In the first place the dissemination may be caused by mineralization along both sides of a fissure, parallel with the schistosity—that is, by the formation of a "bedded vein." Such mineralization is later than metamorphism, and the metamorphic minerals will probably be found to be altered—sericitized, carbonatized, or more rarely silicified.

If, on the other hand, the sulphide minerals were contained in the rock previous to metamorphism, or if they were developed during that process, they will be found intergrown with the metamorphic minerals, such as amphibole, epidote, chlorite, garnet, and albite, and are usually accompanied by some magnetite or ilmenite.

Larger pyritic masses of this kind are, in most cases, probably original products of magmatic concentration; or they may be old fissure veins or replacement veins which have been rendered unrecognizable by deformation; or, finally, they may be of contact-metamorphic origin.

Sparser disseminations, often following certain lines along the strike of the schist, are often called "fahlbands" (the German "fahl" meaning rusty brown and referring to the oxidized outcrops). Such fahlbands, first noted in Kongsberg, Norway,[1] where they enrich the silver veins, may be several miles long and vary in thickness between a fraction of a foot and several hundred

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[1] C. A. Münster, ref. in *Zeitschr. prakt. Geol.*, 1896, p. 93.
J. H. L. Vogt, *idem*, 1899, pp. 177–181.

feet. The enclosing rocks vary from gneiss to mica schist, diorite, and amphibolite. The ore minerals are pyrite, pyrrhotite, zinc blende, chalcopyrite, molybdenite, and sometimes cobalt minerals. They are often intergrown with amphibole or garnet. The fahlbands are rarely of economic importance, but many of them characteristically enrich intersecting veins, causing native silver and gold as well as cobalt and nickel ores to appear at the intersections. This is probably only a special case of the general law that veins are enriched where they cut across belts of pyritic impregnation. Fahlbands rich in cobaltite, with pyrite, chalcopyrite, pyrrhotite, and molybdenite, were worked at Skutterud and Snarum, in the Modum parish, Norway, from 1776 to 1899. For a long time these deposits were among the principal sources of cobalt oxide, which is used to impart a deep blue color to glass and porcelain. According to the older literature quoted by Stelzner and Bergeat[1] the fahlbands at Skutterud lie between gneiss or quartz schist and amphibolite. Other minerals mentioned are malacolite, antophyllite, and rarely graphite and tourmaline. The ores were poor, containing, even when sorted, less than 1 per cent. cobalt. A parallel belt at Snarum is said to be enclosed in amphibolite and contains more copper.

The fahlbands have been variously interpreted. At a time when the crystalline schists were generally considered as altered sediments, they were held to be sedimentary deposits. Dall and Kjerulf,[2] in 1880, held them to be impregnations related to gabbro intrusions. Vogt considered the gray gneiss of Kongsberg as a pressed granite and held that it had been impregnated with sulphides at the same time as the surrounding schists.

That the dissemination of sulphides in its present form is dependent upon dynamo-chemical metamorphism is clearly shown by the minerals with which the sulphides are now intergrown. Sulphide emanations from intrusive magmas at a considerable distance from their source do not usually crystallize with amphibole, pyroxene, and garnet, but rather with calcite, sericite, and quartz as gangue minerals. Still, the recrystallization under pressure does not necessarily explain the ultimate origin of the minerals and it is probably hopeless to speculate on this subject until the metamorphic series at the location of typical

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[1] Die Erzlagerstätten, 1, 1904, pp. 269–271.

[2] Die Geologie des südlichen und mittleren Norwegens, 1880.

fahlbands has been more carefully examined as to the original character of its rocks.

Somewhat similar fahlbands in amphibolite and gneiss are reported in the older literature from Schladming, in Styria, where they enrich intersecting cobalt-nickel veins, and from Les Challanches, in France, where similar relations exist.[1] Recent descriptions from both places show that the so-called fahlbands are in reality narrow veins accompanied by alteration of the wall rocks.[2]

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[1] Stelzner and Bergeat, Die Erzlagerstätten, 1, 1904, pp. 268–269.

[2] C. Schmidt and J. H. Verloop (Schladming), *Zeitschr. prakt. Geol.*, vol. 17, 1909, pp. 271–276.

T. A. Rickard (Challanches), *Trans.* Am. Inst. Min. Eng., vol. 24, 1894, pp. 689–705.

CHAPTER XXII

DEPOSITS OF NATIVE COPPER WITH ZEOLITES IN BASIC LAVAS

GENERAL STATEMENT

Native copper, chalcocite, bornite, much more rarely chalcopyrite, and their products of oxidation are often found in flows of basic lavas, particularly in basalts, associated with minerals of the zeolite group, such as analcite, natrolite, stilbite, chabazite, and laumontite, and the minerals prehnite and datolite; together with these calcite, quartz, chalcedony, chlorite, epidote, and adularia may be present, sometimes in predominating quantity. These gangue minerals, together with the copper minerals, fill vacuoles or blowholes in the basic rocks or replace the rock. Pyrite and sulphides of metals other than copper rarely occur.

These deposits have been formed near the surface under conditions which will be discussed in a following paragraph. The mineral association does not indicate a deep-seated origin.

Instances of native copper occurring in this manner are plentiful, though the occurrences are not always of economic importance. Among the numerous localities the following may be mentioned: The Faeroer,[1] north of Scotland; Sterling,[2] in Scotland; Oberstein a. d. Nahe, Germany; Sao Paulo,[3] Brazil; the Kristiania region,[4] Norway; the Triassic "traps" of New Jersey[5] and Connecticut; New Guinea;[6] the Transbaikalian provinces[7] on the Dochida River; the Bay of Fundy,[8] Nova Scotia.

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[1] F. Cornu, *Zeitschr. prakt. Geol.*, vol. 15, 1907, p. 321.

[2] Carl Hintze, Handbuch der Mineralogie, 1898.

[3] E. Hussak, *Centralblatt f. Min.*, 1906, p. 333. (No zeolites; copper between the peripheral covering of the amygdules, consisting of an iron silicate and the filling of chalcedony.)

[4] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, vol. 7, 1899, p. 12.

[5] Volney Lewis, *Ann. Rept.*, Geol. Survey New Jersey, 1907, pp. 157 and 165.

[6] R. Beck, Lehre von den Erzlagerstätten, I, 1909, p. 345.

[7] *Idem*, p. 346; also *Zeitschr. prakt. Geol.*, vol. 9, 1901, p. 391. (With opal, chalcedony, calcite, epidote, and prehnite.)

[8] R. W. Ells, Copper in the provinces of Nova Scotia, New Brunswick, and Quebec, *Min. Res. Can.*, Geol. Survey Canada, 1904, 58 pp.

The first six of these occurrences have not been worked as copper deposits. The last four are of some economic importance.

A pre-Cambrian series of basaltic rocks in the Lake Superior region contains the most prominent example of this class of deposits in the world, which will be described in more detail below.

In eastern Oregon,[1] about 20 miles east of Baker City, and along the Snake River cañon, opposite the Seven Devils Mountains in Idaho, are extensive areas covered by a basaltic amygdaloid flow of Jurassic or Triassic age. This rock contains native copper and chalcocite, sparsely disseminated or along obscure fracture zones, in association with epidote, chlorite, calcite, and zeolites. The ores are of low grade and have not yet been worked with profit.

Another occurrence of interest in the White River region in Alaska has recently been described by Adolph Knopf.[2] The basaltic amygdaloids, with tuffs and breccias, are interbedded with sediments of Carboniferous age and have probably been erupted in part under submarine conditions. Placer copper is common in the creeks, and some large masses have been found. The copper minerals are chalcocite, chalcopyrite, and native metal in stringers and seams, with prehnite, laumontite, thomsonite, and calcite; also native copper with zeolites filling blowholes in reddish, highly amygdaloid lava.

These statements will serve to show that the zeolitic copper deposits in basaltic lavas represent a type of world-wide distribution; the same processes of concentration are evidently applicable to all cases.

ORIGIN OF THE ZEOLITIC COPPER ORES

Probable Source of Copper.—Basic igneous rocks such as gabbro, diabase, basalt, some andesites, and basaltic flows designated melaphyres or amygdaloids probably always contain copper, in some cases as much as 0.1 or 0.2 per cent., more commonly about 0.02 per cent. of the metal (p. 8). According to Volney Lewis and F. F. Grout, the copper is present as a silicate, possibly in part as a sulphide, such as bornite or chalcocite.

It is likely that the copper is present mainly as a silicate

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[1] W. Lindgren, The gold belt of the Blue Mountains of Oregon, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, pp. 551–776.

[2] *Econ. Geol.*, vol. 5, 1910, pp. 247–256.

in effusive rocks, while in intrusive rocks a part of the copper is held as a sulphide. Sulphur compounds escape in large quantities from basalts during eruption. In the intrusive rocks this sulphur is retained.

The minute quantity of copper in these rocks may become concentrated to valuable deposits in various ways. This can be effected at a certain depth below the surface by circulating waters of atmospheric origin, or by ascending currents of thermal waters of deep-seated origin, or during regional metamorphism in the zone of combined fracture and flow. It is believed that any of these processes may result in copper deposits, few of which, however, will be extensive or valuable.

The concentration of the copper stands in intimate connection with the development of zeolites, and it will first be necessary to discuss this process.

The Occurrence of Zeolites and the Process of Zeolitization.—The zeolites are mainly aluminum-calcium silicates with 8 to 15 per cent. of water of hydration. Sodium or sometimes potassium may replace part of the calcium, and in some zeolites barium or strontium is present. Magnesium does not usually enter into their composition, but appears in the associated chloritic minerals.

Analcite, a sodium-aluminum silicate containing 8 per cent. H_2O, is also considered to belong to the zeolites. Prehnite, $H_2Ca_2Al_2Si_3O_{12}$, containing 4.37 per cent. H_2O, and datolite, $H_2Ca_2B_2Si_2O_{10}$, with 5.63 per cent. H_2O, do not strictly belong to the zeolites, but are commonly associated with them. The most common zeolites are natrolite, desmine, chabazite, apophyllite, thomsonite, and laumonite.

The zeolites can be easily produced by synthesis at temperatures of 100° to 500° C. Some of them, like apophyllite, are soluble in water with or without CO_2 or Na_2CO_3 at 180° to 189° C. at a pressure of 10 to 12 atmospheres, crystallizing again after cooling. Chabazite was recrystallized in a closed tube by Doelter at 150° C., also in fluid CO_2 at 30° C. Datolite and prehnite have not been produced by synthesis.

The different modes of occurrence of zeolites may be classified as follows:[1]

1. Filling amygdules and veins in flows of basic lavas. This is the most common occurrence.

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[1] References generally from Hintze's "Mineralogy."

2. Filling miarolitic cavities in granite and here probably formed shortly after the consolidation of the rock.

3. In pegmatite dikes as products of the last epoch of crystallization.[1]

4. As veins or coatings of joint planes in granite or gneiss or various volcanic rocks; here associated with calcite and sometimes with amethystine quartz, occasionally with albite.

5. In contact-metamorphic mineral deposits in limestone.

6. In the so-called Alpine type of veins, common in Switzerland, Tyrol, and the French Alps. With quartz, adularia, and many rare and well-crystallized minerals.

7. In mineral veins, associated with sulphides. Very rare and mainly as last products of crystallization. Andreasberg, Germany; Kongsberg, Norway; Guanajuato, Mexico; Arqueros and Rodaito, Chile.

8. As products of deposition of hot springs at their orifices, as at Plombières, Bourbonne-les-Bains and Luxeuil in France; at Oran, Algeria; at Hunter and Boulder Hot Springs, Montana.

9. In deep-sea deposits. Phillipsite has frequently been found in the mud brought up by the dredges.

Undoubtedly zeolites may form at low temperatures, as shown by the mentioned occurrence of phillipsite. The ranges of stability of the various zeolites may differ considerably. C. Doelter[2] believes that the limits for the development of analcite lie between 180° and 440° C.; for natrolite he thinks they are considerably lower, say from 0° up to 180° C. Experimentally the latter mineral has been obtained at a temperature as low as 90° C. It is probable that datolite, prehnite, and adularia do not develop at temperatures much lower than 100° C.

In many occurrences it can be shown that the zeolites formed as the last phase of the consolidation of a magma; their mode of appearance in pegmatite dikes and close to igneous contacts points plainly to this origin.

Furthermore, they seem to require stagnant, quiet conditions, such as prevailed in cooling bodies or in rocks impregnated with warm water, as in the Roman brickwork at Plombières. Their general absence from mineral veins shows that swiftly moving or ascending water is distinctly unfavorable for their development.

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[1] W. C. Brögger, *Zeitschr. Kryst. Min.*, Bd. 16.

[2] C. Doelter, Minerogenese und Stabilitätsfelder der Minerale, *Tsch. M. und. P. Mitt.*, vol. 25, 1906, pp. 79–112.

The accepted authorities are more or less vague in their statements as to the formation of zeolites, especially in amygdaloid rocks. The most common statement is that these minerals are deposited by percolating waters. Van Hise considers them to be formed in the zone of cementation by descending surface waters, and also by similar waters percolating through lava flows and extracting the material for the zeolitization from the rock itself.[1]

Zeolites are manifestly unstable in the zone of weathering and must have been formed at some depth. Of late years the opinion has been gaining ground[2] that zeolitization, in basic volcanic rocks, is distinctly connected with the cooling processes and in fact should be regarded as an after-effect of volcanism, their deposition taking place in the still hot rocks.

That zeolitization is far from being simply an effect of the leaching by surface waters is shown by the absence of the zeolites from large areas of basic flows, many of them full of vacuoles or blowholes. Few occurrences have been recorded from the Hawaiian flows, which are apparently well suited for their deposition, nor from the extensive flows of the Columbia lava in Oregon and Washington.[3] There are, therefore, certain conditions—not yet fully elucidated—which are necessary for the deposition of zeolites. It is probable that their development would be greatly furthered if the eruption of the effusive rock took place under water; the sea water would cool the surface of the flow and a slow downward movement would be caused in the porous rock. Besides, these conditions would give rise to a system, cool at one end, hot at the other, in which circulation competent to effect concentration would be initiated.

One of the most convincing proofs that zeolitization follows closely upon eruption has been given by Knopf,[4] who describes an occurrence in the White River region of Alaska where a sheet of amygdaloid rock containing copper is covered by a coarse

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[1] C. R. Van Hise, *Mon.* 47, U. S. Geol. Survey, 1904, pp. 333 and 633.

[2] J. Volney Lewis, *Ann. Rept.*, State Geol. New Jersey, 1907, p. 167.
Alfred Harker, The natural history of igneous rocks, 1909, p. 308.
Adolph Knopf, *Econ. Geol.*, vol. 5, 1910, pp. 247–256.
C. N. Fenner, The Watchung basalt and the paragenesis of its zeolites, *Annals.*, New York Acad. Sci., vol. 20, pt. 2, 1910, pp. 97–187.

[3] According to F. C. Calkins (oral communication) zeolites were found at one locality in these lavas in the John Day region.

[4] Adolph Knopf, *op. cit.*, p. 251.

pyroclastic bed, the breccias of which include fragments of the cupriferous amygdaloid, proving that the filling of the vacuoles took place during the interval between successive extrusions of lava. In places—for instance, where the cupriferous zeolites occur in fissures—there was probably a longer interval, but all the infilling was probably accomplished before the rock had cooled.

C. N. Fenner has recently investigated the zeolites of certain Triassic basalts of New Jersey, which cover land sediments and old playas or shallow desert basins of the same age, and finds that the zeolitization took place mainly where the basalt flows covered the shallow lakes; he concludes that the circulation originated from the waters of these lakes. The general process, he says, was that of a slow cooling of the igneous rock, through which aqueous solutions were percolating. Material for solution was contributed by the basalt and by the previously evolved sublimates. The character of the minerals changed during the cooling. Pyrite and chalcopyrite are among the metallic minerals; native copper is not mentioned, but occurs at many places in these Triassic flows. Three periods of crystallization are distinguished. Beginning with the oldest they are as follows:

1. Boric acid period	(*a*)	Albite, quartz, garnet, amphibole, specularite, sulphides.
	(*b*)	Datolite, prehnite, pectolite, amphibole, specularite, sulphides.
2. Zeolite period		Andesine, chabazite, heulandite, stilbite, natrolite, laumontite, apophyllite, amphibole, chlorite, specularite, sulphides.
3. Calcite period		Thaumasite, calcite, gypsum, amphibole, chlorite, specularite, sulphides.

This combination is of special interest, as it shows a peculiar combination of high-temperature minerals like garnet and amphibole with the zeolitic deposits. Extensive replacements were noted, similar to processes described long ago by Pumpelly, from observations in the Lake Superior copper mines. Minerals stable under new conditions replace those formed in older crystallizations. Datolite, prehnite, pectolite, chabazite, stilbite, natrolite, apophyllite, and calcite—all replace the older albite. Quartz is replaced by calcite and various zeolites. Datolite is replaced by zeolites.

Knopf[1] has justly stated that "any theory accounting satis-

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[1] Adolph Knopf, *op. cit.*, p. 253.

factorily for the zeolites will also account for the copper." There is surely just as much of a concentrating process involved in obtaining fluorine for apophyllite, boron for datolite, or barium for harmotome as there is in producing an ore with 2 per cent. copper from an amygdaloid containing 0.02 per cent. of the metal.

Following Lane (p. 438), I believe that the water of seas or lakes, mingling with the exhalations from the magma, decomposed the copper silicate contained in the pyroxenes, and that the resulting chlorides of iron and copper were decomposed by silicates or carbonates of calcium, with the formation of native copper, ferric oxide, and calcium chloride.

THE LAKE SUPERIOR COPPER DEPOSITS

In the following list of references on the copper deposits of the Lake Superior region only the more important works of the extensive literature are mentioned.

H. Credner, *Neues Jahrb.*, 1869, pp. 1–14.

R. Pumpelly, Geol. Survey Michigan, vol. 1, pt. 2, 1873.

R. Pumpelly, The metasomatic development of the copper-bearing rocks of Lake Superior, *Proc.*, Am. Acad. Arts and Sci., vol. 13, 1877–1878, p. 253.

C. Rominger, Geol. Survey Michigan, vol. 5, 1895.

R. D. Irving, The copper-bearing rocks of Lake Superior, *Mon.* 5 U. S. Geol. Survey, 1883.

H. L. Smyth, Theory of origin of the copper ores of the Lake Superior district, Science, vol. 3, 1896, p. 251.

M. E. Wadsworth, Origin and mode of occurrence of the Lake Superior copper deposits, *Trans.*, Am. Inst. Min. Eng., vol. 27, 1898, pp. 669–696.

L. Hubbard, Keweenaw Point, Geol. Survey Michigan, vol. 6, pt. 2, 1898.

A. C. Lane, Geological report on Isle Royale, Geol. Survey Michigan, vol. 6, pt. 1, 1898.

A. C. Lane, The theory of copper deposition, *Ann. Rept.*, State Geol. Michigan, 1903.

A. C. Lane, Native copper deposits, *Bull.* 13, Canadian Min. Inst., February, 1911, pp. 81–87.

A. C. Lane, The Keweenaw series of Michigan, Mich. Geol. and Biol. Survey, Lansing, 1911.

T. A. Rickard, The copper mines of Lake Superior, 1905, p. 164. (Excellent brief description of geology and technology.)

L. C. Graton, *Mineral Resources*, U. S. Geol. Survey, Annual publication, 1906–1907.

B. S. Butler, *idem*, 1908–1917.

C. R. Van Hise and C. K. Leith, *Mon.* 52, U. S. Geol. Survey, 1911, pp. 578-592.

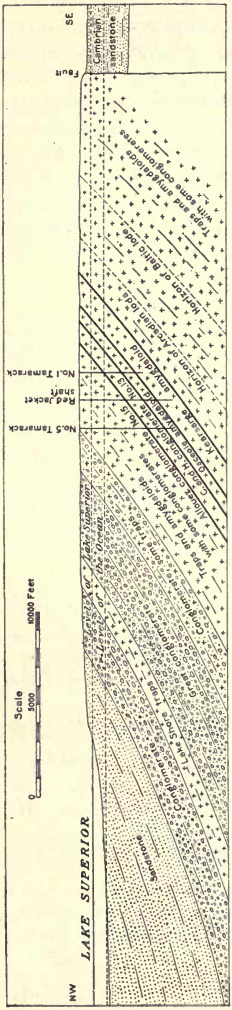

FIG. 132.—Section across the copper-bearing beds at Houghton, Michigan. *After A. C. Lane.*

General Occurrence.—The great deposits of native copper in the Keweenawan volcanic flows and conglomerates of the pre-Cambrian in Michigan form one of the principal items of the copper wealth of the United States. These deposits are mainly concentrated in Houghton County, on the Keweenaw peninsula of northwestern Michigan, on the southern shore of Lake Superior (Fig. 116). The copper belt continues, however, in a northeasterly direction to the point of the peninsula and southward into Ontonagon County and into northern Wisconsin and eastern Minnesota, having a total length of about 300 miles. The cupriferous formation is found also on Isle Royale and at Michipicoten in Ontario, Canada, opposite Keweenaw Point. The present productive belt extends in a northeasterly direction from the Lake mine, 43 miles southwest of Houghton, the center of the industry, to the Cliff mine, 22 miles northeast of that place.

The so-called Keweenawan series, the uppermost part of the Algonkian, unconformably covers the Huronian, which in turn rests discordantly on the Archean greenstones and gneisses.

The Keweenawan series forms a huge synclinorium, bounding and underlying the western part of Lake Superior. The upper part of the series consists of 10,000 or 12,000 feet of red arkose sandstones and shales; the lower part consists of a vast accumulation of basaltic lavas perhaps 25,000 feet in total thickness. On the Keweenaw peninsula the whole series strikes northeast,

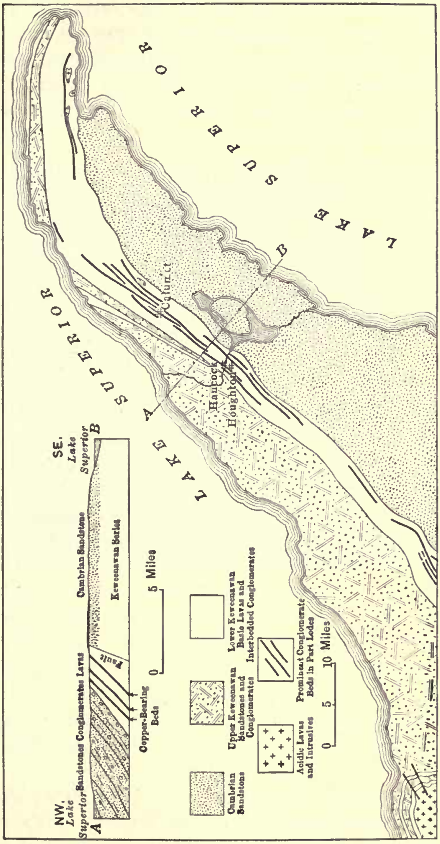

FIG. 133.—The Michigan copper region. *After Seaman and Lane.*

parallel to the direction of the peninsula, and dips northwest from 30° to 75°. The sandstones follow the northwest coast. The rapidly alternating series of compact diabases or basalts (traps) and amygdaloid beds occupies the central belt. On the southeast the Keweenawan series is cut off by a long fault line parallel to the strike, and the southeast coast of the peninsula is underlain by horizontal non-productive Cambrian sandstones. Embedded in the volcanic flows are a few strata of sandstone, shale, or conglomerate. Quartzose porphyries (felsites) erupted in the same epoch enter conspicuously into the composition of these conglomerates (Figs. 132 and 133).

Copper in visible grains is widely distributed through the amygdaloid rocks, but is of economic importance in only a few places. The deposits which now yield the great bulk of the production are beds of amygdaloid rocks of great persistency. Much copper is also mined from a bed of volcanic conglomerate, called the Calumet conglomerate. Veins cutting across the strike of the beds were mined in the early years, but are now of little importance. The mining operations were begun about 1846.

The Calumet Conglomerate.—This bed is worked mainly in the Calumet & Hecla and Tamarack mines; in the former mine it is opened for a distance of nearly 2 miles along the strike by 10 incline shafts and one vertical shaft, with an aggregate of about 200 miles of workings. The Red Jacket vertical shaft is 4,900 feet deep and reaches to the fifty-seventh level of the incline. The Tamarack mine, in which the continuation of the shoot in depth has been found, reaches the conglomerate by four vertical shafts from 3,409 to 5,309 feet in depth, the latter being one of the deepest shafts in the world. Only parts of the conglomerate are of profitable grade, the ore-shoot trending north on the bed. In depth it appears leaner and less regular in tenor. Along the surface only parts of the bed contain commercial ore; it is poor both northeast and southwest of the Calumet shoot. The tenor of the ore handled has decreased from 4 per cent, of copper in former years to 1.5 per cent. in 1916. The ore itself is undoubtedly becoming leaner, but the apparent decrease is to a considerable extent due to improvements in mining and milling that allow the handling of lower grades of ore.

The conglomerate is 10 to 25 feet thick, dips 36° to 39° northwest, and forms a compact reddish-brown rock easily breaking across the pebbles. The copper occurs mainly as small particles in the cement between the cobbles, which are well rounded and consist of quartz porphyry with some basic igneous rocks. The hanging wall is a dark fine-grained diabase, the footwall a thin layer of sandstone. About one-third of the copper production of Michigan is obtained from this conglomerate.

The Amygdaloids.—The amygdaloid copper-bearing beds, which occur at seven principal horizons, are named the Baltic, Kearsarge, Pewabic, Osceola, Isle Royale, Atlantic, and Winona amygdaloids and are worked by a dozen large mines. These beds are vesicular basalts, usually brownish in color, with earthy fracture and filled with amygdules of calcite, epidote, and zeolites (Fig. 134). The copper occurs in these, but also replaces the rock itself Some native silver, in places intergrown with copper, occurs in the amygdaloids; scarcely any is found in the conglomerate. The flows are naturally more vesicular in the upper part than in the bottom part. Both width and distribution of copper are irregular. The Osceola bed is worked by the Calumet & Hecla, Tamarack, and Osceola mines; in the Osceola it is developed to a depth of 4,500 feet on the incline. In the Calumet & Hecla mine this amygdaloid is 30 to 35 feet thick, but the mineralization is mainly confined to a strip 8 or 10 feet thick along the hanging wall and also a streak along the footwall. In the Baltic mine the amygdaloid bed is from 15 to 80 feet in

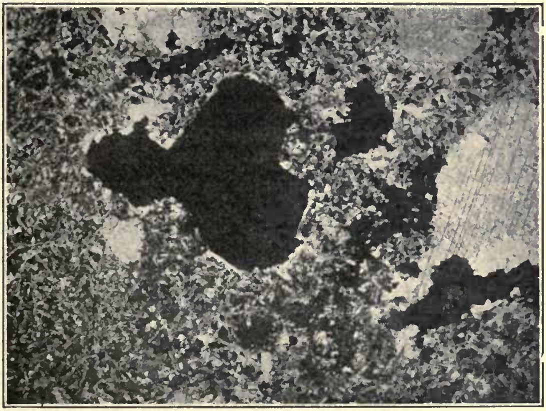

FIG. 134.—Amygdaloid basalt, Houghton, Michigan. Black areas, native copper; larger areas represent fillings of blow holes, with calcite at right margin. Smaller black areas represent replacement of igneous rock by copper. Magnified 15 diameters.

stoping width; this deposit has produced much coarse copper and contains veinlets of chalcocite, bornite, and copper arsenides. The Kearsarge amygdaloid is worked continuously for a distance of 12 miles.

The amygdaloid ores yield an average recovery of 0.88 per cent. copper, actually varying from 0.5 to 1.6 per cent. In 1916, 716,640 ounces of silver was obtained, in small part as nuggets or "pickings," but mainly from the electrolytic refining process, to which a part of the copper produced was subjected. The largest lump of silver on record from the district weighed 12 pounds and was found in the Mass mine, in the southern part of the district.[1] The average recovery of silver per ton of ore is 0.33 ounce per ton.

The Veins.—A third mode of occurrence of the copper ore is as veins following fracture zones, in the northern part of the peninsula and in Ontonagon County to the south. During the early years of the industry these veins yielded much copper, but are at present of little importance. Most of the veins cross the bedding and stand at a steep angle, though in Ontonagon County many strike veins are also found. In places they are also parallel to the dip of the strata. Some of them could be followed by the drift for a distance of 2,000 or 3,000 feet. According to Credner, Pumpelly, and Irving many of them were wide, though they averaged only 3 feet.

In part these veins were formed by filling, but they were chiefly the result of metasomatic replacement. Much of the native copper was coarse; some masses of unusual size were found, the most famous being that encountered in the Minnesota vein in 1880. The mass weighed 500 tons, was 46 feet long, 18.6 feet wide, and 8.5 feet thick.[1] At the Cliff mine many masses which weighed from 40 to 100 tons were discovered. From the vein the copper seems to have had a tendency to extend into the various amygdaloid flows. Most of the veins became impoverished at a depth of a few hundred feet.

The amygdaloid beds are cut by many minor cross fractures and slip faults, but according to the accounts these contain little or no copper.

Mineral Association.—In all three modes of occurrence the mineral association is the same. It consists of native copper, quartz, calcite, chlorite, epidote, datolite and prehnite, with a

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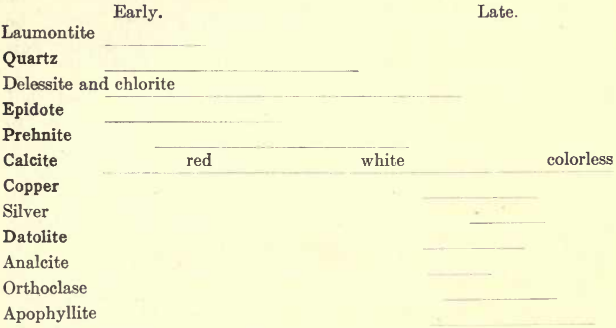
[1] T. A. Rickard, *op. cit.*

number of zeolites. There is always some ferric oxide, in places staining the ore deep red or brown. Chalcocite and some rare arsenides of copper are entirely subordinate. The following table, adapted by Lane from Pumpelly, gives the general paragenesis. The more common minerals are printed in heavy type.



The stages of alteration and filling in the amygdaloid rock are indicated as follows by Pumpelly: (1) Decomposition of the ferromagnesian silicate and deposition of iron-rich chlorite (delessite). (2) Individualization of the non-alkaline silicates (laumontite, prehnite, and epidote). (3) Deposition of quartz. (4) Introduction of native copper, with replacement of prehnite by delessite. (5) Appearance of the alkaline silicates (analcite, apophyllite, adularia), representing the decomposition of labradorite in the original rock. Many interesting replacements have taken place in the rock itself: Prehnite is pseudomorphic after labradorite and many amygdaloids are largely prehnitized. This prehnite may in turn be replaced by adularia and the latter may change into epidote and quartz. Sericite is absent. Needles of actinolite are sometimes seen in the amygdules. Datolite is present in flinty, massive and crystallized form.[1]

Origin.—The ore deposits are of considerable antiquity and it is probable that the present mine waters have little to do with

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[1] An unusual occurrence of copper is in the Nonesuch sandstone of Ontonagon County where it forms replacement of the cement. Irving, as well as Van Hise and Leith, states that the copper contains cores of magnetite. K. Nishio found that the "magnetite" consisted of a black hydrocarbon. *Econ. Geol.*, vol. 14, No. 3, 1919.

the origin, though they may have effected slight changes and local concentration. U. S. Grant[1] has shown that the deposits were in existence when the Cambrian strata along the great fault in the southeastern part of the peninsula sank to the level of the Keweenawan series; it is indeed most likely that they were formed before the deposition of this Cambrian sandstone. The continental Quaternary ice sheet doubtless swept away the altered upper part of the beds, so that the native copper now outcrops almost at the surface.

The association of minerals is entirely different from that found in ordinary fissure veins, in which, we have reason to believe, the deposition was effected by ascending thermal solutions. The so-called "Alpine veins" (p. 631) offer some analogies; likewise the veins of Andreasberg and Kongsberg (p. 623). There is also some resemblance to propylitization, but in that process zeolites rarely form.

Evidence has already been adduced that all the fresh diabasic and basaltic rocks of the series contain copper, probably as a silicate, and throughout the vast extent of the Keweenawan the amygdaloids show traces of the metal itself.

Van Hise[2] says:

> There is scarcely a locality in the Lake Superior region where the Keweenawan basic lavas occur in which small amounts of copper are not found. Almost every porous amygdaloid shows flakes of it. * * * To me, the almost universal association of small quantities of copper with the Keweenawan lavas is the most conclusive evidence that these lavas are the source of the metal.

Pumpelly suggested that the presence of ferric oxide and epidote (in which the iron is in the ferric state) indicated a reduction of copper salts (sulphate and carbonate) by the ferrous minerals abundantly present in the rock.

Lane[3] has recently proposed a modification of this view, based on some valuable experiments undertaken by G. Fernekes. After the submarine effusion of the lavas, sea water penetrated

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[1] *Bull.* 6, Wisconsin Geol. and Nat. Hist. Surv., 1901.

[2] C. R. Van Hise, *Mon.* 47, U. S. Geol. Survey, 1904, p. 1103.

[3] A. C. Lane, Salt water in the Lake mines, *Proc.*, Lake Superior Min Inst., vol. 12, 1906.

A. C. Lane, Native copper deposits, *Jour.*, Canadian Min. Inst., vol. 14, 1911, pp. 316–325.

the beds, decomposing the silicates and converting a part of the iron and all of the copper to chlorides. The reduction of the cuprous chloride was effected by calcium salts, with the formation of ferric oxide and calcium chloride. The process may have persisted during the slow cooling until fissures and joints had formed in the beds, and this would explain the deposition on such fractures.

The experiments of Fernekes[1] have shown that metallic copper is precipitated, together with ferric oxide, from a mixture of ferrous and cuprous chlorides, in a tube one end of which is heated to 200° to 280° C., while the other end is cooled. The precipitation takes place, however, only in the presence of a substance or mineral which neutralizes the hydrochloric acid, hydrolyzed from $FeCl_3$. Calcium carbonate, datolite, and prehnite were found to have this neutralizing property. No results were obtained with laumontite and labradorite. The equations are:

$$2FeCl_2+2CuCl_2=Cu_2Cl_2+2FeCl_3.$$
$$2FeCl_2+2CuCl=2Cu+2FeCl_3.$$

The presence of silver is explained by the solubility of the chloride of that metal in strong salt solutions. Lane expresses the above equation schematically as follows:

$$2FeCl_2+2CuCl+3CaSiO_3=2Cu+Fe_2O_3+3SiO_2+3CaCl_2.$$

Owing to the strong dehydrating power of chloride solutions, ferric oxide will be deposited instead of limonite. At an earlier date H. N. Stokes[2] had ascertained that hornblende and siderite precipitate native copper from sulphate solution at 200° C., under conditions similar to those in Fernekes's experiments.

The boron and fluorine in datolite and apophyllite were probably also concentrated from the amygdaloids. Much of the carbon dioxide and chlorine may well have been contributed by the volcanic flow itself. Copper and silver form an alloy at 540° C.[3] As the two metals exist in close contact in the Lake Superior deposits the conclusion is justified that these deposits were formed at a lower temperature.

Whitney, Pumpelly, and Wadsworth, have advocated a theory of deposition by descending surface waters. Pumpelly assumed

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[1] G. Fernekes, *Econ. Geol.*, vol. 2, 1907, pp. 580–584.

[2] *Econ. Geol.*, vol. 1, 1906, p. 648.

[3] F. E. Wright, Science, vol. 25, 1907, p. 389

that the copper sulphide present in the beds was first oxidized to sulphate and carbonate and subsequently reduced, but, as has been shown, this hypothesis is not necessary. Van Hise and H. L. Smyth believed the deposits to be caused by ascending thermal waters, but the whole character of the mineralization is directly opposed to such a view.

Mine Waters.[1]—The present condition of the underground waters in the copper region is most interesting. Lane has shown that the water in the upper levels is soft and potable and has the normal composition of surface waters. It decreases in quantity as depth is gained and ceases at a depth of 1,000 to 1,500 feet below the surface.

ANALYSIS OF NORMAL SURFACE WATER FROM MICHIGAN COPPER DISTRICT

(Parts per million)

Ca	19	SiO_2	10
Mg	4	CO_3	40
Na	2.3	SO_4	6
Cl	3.5	$(Al,Fe)_2O_3$	1.5
			86.3

As depth is attained in the mines the quantity of chlorine and calcium increases very materially, and at the same time the mine water is less abundant. Finally, at a depth of 3,000 to 5,000 feet, the mine waters are almost entirely absent; they constitute feeble drips here and there and, of course, may collect in small quantities in the sumps. They are extremely strong solutions of calcium chloride with bromine and many other substances in small quantities.

Other samples of these waters contain an appreciable amount of zinc and some strontium.

In the most concentrated waters 99 per cent. of the salts consist of the chlorides of calcium and sodium, and three-fourths of the remainder is sodium bromide.

Lane points out that waters of this composition are not unknown in other deep sedimentary series and suggests that they may be "connate" waters—that is, residual waters from those deposited with the sediments and derived from the Keweenawan pre-Cambrian sea. The strong percentage of bromine is

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[1] A. C. Lane, Salt water in the Lake Mines, *Proc.*, Lake Superior Min. Inst., vol. 12, 1906, pp. 154–163.

additional evidence that we have here really to deal with a residual sea water, a remnant of that which long ago was active in forming these deposits. Van Hise and Leith believe that the deep mine water may represent the residuum of the ore forming solutions but do not consider it as residual sea water.

ANALYSIS OF MINE WATER, QUINCY MINE

From drippings on 55th level north of No. 6 shaft. G. Fernekes, analyst. Depth about 4,000 feet[1]

(Grams per liter)

Cl	176.027	SiO_2	.020
Br	2.200	$(Fe,Al)_2O_3$	.010
Ca	86.478	Mn	.004
Na	15.188	Cu	.016
K	.411	Ni	.006
SO_4	.110	Mg	.020
			280.490

Total solids determined 280.500.
Traces of boron and strontium. No barium, lithium, or carbon dioxide.

Rock Alteration.—The ores have often a yellowish green color from disseminated epidote; in places they are bleached and contain much calcite and chlorite. Van Hise and Leith[2] have published several analyses of such rocks and Lane has examined bleached pebbles in the Calumet conglomerate. These analyses show that the alteration lacks uniformity; they do not indicate the influence of normal weathering nor are the changes similar to those of hydrothermal alteration of wall rocks of veins. The silica has decreased in some rocks; in others, there is strong enrichment of lime or soda according to the stage of mineral paragenesis reached. There is no concentration of potash. The rock is moderately hydrated but there is no kaolin present. The results are perhaps most similar to the widespread "propylitic" alteration of igneous rocks in hydrothermal areas.

Mining and Smelting Operations.—In the copper mines of Lake Superior mining operations are conducted on a large scale. The total amount of copper ore (locally called "rock") hoisted in Michigan per annum is about 12,000,000 tons. This is crushed coarse with steam stamps, each one having a daily capacity of

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[1] At the Franklin mine the limit between the upper potable waters and the salt waters is about 1,300 feet below the surface, or 200 feet below sea level. In general the chloride waters appear about sea level.

[2] *Mon* 52, U. S. Geol. Survey, 1911, p. 583.

500 to 700 tons; wet concentration is used with jigs, tables, etc., the resulting concentrates (locally called "mineral") amounting in 1916 to about 200,000 tons, of an average copper content of 66 per cent.

This concentrate of native copper is smelted and refined by a single operation in reverberatory furnaces, the smelting works being located in the district and at Buffalo. A small part of the copper is electrolytically refined in order to eliminate the small amount of arsenic contained. A demand for copper containing arsenic that has recently arisen has resulted in a decrease of the quantity refined by the electrolytic process. The annual copper production of Michigan increased steadily to 1905 when it reached 230,000,000 pounds. Since then the changes have not been great, except that re-working of tailings has lately increased the output. It was roundly 256,000,000 pounds in 1917. The reserves of amygdaloid copper-bearing rock are of great extent.

THE COPPER DEPOSIT OF MONTE CATINI

The celebrated copper deposit of Monte Catini, on the western coast of Italy, near Livorno (Leghorn) and the ancient Etruscan city of Volterra, has been described by many authors. A. Bergeat has given an excellent review of this literature, in connection with his own observations.[1] Another detailed description is given by L. de Launay.[2] The mines have been worked to a depth of 850 feet.

Irregular laccolithic stocks of diabase, with some gabbro, break through Eocene marly limestones and siliceous shales; near the contacts the igneous rock is in part glassy, so that the intrusions clearly took place near the surface. The ore occurs exclusively in the diabase, particularly in its lower part at or near the contact, but also reaches the surface. In the ore-body the diabase is crushed to reddish clayey masses seamed with zeolites and calcite. The ores contain native copper in crevices and druses, with calcite, prehnite, datolite, analcite, and laumontite; also sulphides, especially chalcocite, bornite, and chalcopyrite, sometimes massive, but partly in large and small rounded con-

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[1] A. W. Stelzner and A. Bergeat, Die Erzlagerstätten, vol. 2, 1906, pp. 835–842.

[2] Métallogénie de l'Italie, *Compte rendu*, 10th Internat. Geol. Congress, vol. 1, 1907, pp. 603–621.

cretions surrounded by clayey, crushed rock and consisting of the several sulphides in concentric intergrowth. The tenor and distribution of the ores are very irregular.

The whole aspect of this unique deposit seems to indicate that the copper was concentrated from the diabase shortly after its consolidation and the crushing which followed. It is more difficult to point to the source of the concentrating waters, but it is probably safe to say that the present ground waters have had nothing to do with the formation of the ore.

There are remarkable similarities between the mineral association at Monte Catini and that of the amygdaloid flows of the Lake Superior region, and the processes of concentration may, in the main, have been identical.

NATIVE COPPER WITH EPIDOTE IN BASIC LAVAS (CATOCTIN TYPE)

In some copper deposits contained in basic lavas the zeolites are absent and the mineral association is mainly native copper, epidote, quartz, and calcite. Such occurrences, which are of slight economic importance, have been found in the Appalachian region in Virginia and Pennsylvania.[1]

The rocks are basaltic flows of pre-Cambrian age, in part amygdaloid, in part schistose. They contain, in irregular fractures and along shear zones, abundant epidote, native copper, calcite, and chlorite; in places chalcopyrite and bornite occur in the gangue or in the rock itself. Weed named this group of ores the "Catoctin type" and suggested that it owed its origin to infiltration from the present surface. This seems improbable; more likely the copper was extracted from the basic flows shortly after their eruption and consolidation. The derivation of the waters is uncertain; at any rate they were not ascending thermal waters rich in carbon dioxide, for under such influences epidote could hardly be expected to form.

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[1] W. H. Weed, Types of copper deposits in the southern United States, *Trans.* Am. Inst. Min. Eng., vol. 30, 1900, pp. 449–504.

W. C. Phalen, Copper deposits near Luray, Virginia, *Bull.* 285, U. S. Geol. Survey, 1906, pp. 140–143.

G. W. Stose, Copper deposits of South Mountain, Pennsylvania, *Bull.* 430, U. S. Geol. Survey, 1909, pp. 122–131.

CHAPTER XXIII

LEAD AND ZINC DEPOSITS IN SEDIMENTARY ROCKS; ORIGIN INDEPENDENT OF IGNEOUS ACTIVITY

Characteristic Features.—The lead and zinc deposits which form the subject of this chapter represent a type of world-wide distribution and, in spite of local variations, of remarkably constant characteristics. They appear to be entirely independent of igneous rocks and occur in limestones, dolomites, cherts (derived from limestone), or calcareous shales. In the United States this type is represented by the ores in the limestone of the Mississippi Valley; the largest deposits are in Missouri.

The mineral composition is simple, and the ore minerals few. Galena and zinc blende are essential constituents, with their train of oxidized minerals near the surface[1] (sulphates, carbonates, and silicates); there is more or less pyrite, almost always marcasite, occasionally a little chalcopyrite. Gold, antimony, arsenic, and molybdenum are conspicuously absent; in some districts the galena contains a little silver, but on the whole the deposits are non-argentiferous. Cadmium is often contained in the zinc blende, which is mainly red, light brown, or yellow and carries little iron. Cadmium sulphide, greenockite, occurs as a secondary mineral. Nickel and cobalt are often present in small quantities. Among the gangue minerals dolomite is the most characteristic; quartz in crystals is not common, but a secondary chert with bitumen is typical of many districts; barite is found, but is not characteristic.

The ores lie in zones of local brecciation or in crevices (gash veins) or joints which have been enlarged by solution. Less commonly they occupy fault fissures; sometimes they are purely metasomatic, the minerals occurring disseminated in limestone

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[1] The oxidation of lead and zinc sulphides is treated in Chapter XXXI. The principal oxidized zinc minerals are smithsonite and calamine, while hydrozincite and willemite are rarer; goslarite, the soluble sulphate, is frequently found as efflorescences but not in quantities sufficient to be regarded as an important ore mineral.

or dolomite and closely following certain sedimentary horizons. Even in this case they are not spread over irregular areas, but tend to follow certain lines in the plane of stratification (so-called "runs"). In regions of slightly disturbed strata many observers have noted the tendency of the ore to follow pitching troughs. The ores usually lie within a few hundred feet of the surface and are oxidized in the vicinity of the water level. Frequently they are found below impervious shale beds.

Origin.—Simple as the deposits of this type are, the views as to their origin are still divergent. The earliest interpretation of them as marine deposits is generally abandoned; it is recognized that even if the metals are derived from primary ocean sediments the finely divided sulphides must have been concentrated and redeposited. Their epigenetic nature is clear. Some geologists hold the ores to be deposited by ascending waters; others see in them the work of descending surface waters.

In either case American geologists generally believe that atmospheric waters have effected the concentration of the lead and zinc from sedimentary Paleozoic rocks and that igneous agencies have had nothing to do with the deposition. This opinion is not unanimous for a number of investigators have suggested an origin by thermal waters ascending from great depths. Some of these consider that the metals were extracted by the hot waters from the underlying pre-Cambrian rocks, while others believe they can see a relationship of the deposits with deep intrusions and magmatic sources. Beyschlag, Krusch and Vogt in their recently published handbook on ore deposits uphold the theory of origin by thermal waters.

No one can, however, deny that galena and sphalerite are of widespread occurrence in many limestones and dolomites far from regions of deep fissuring and igneous action.[1] Before appealing to igneous agencies it will be advisable to examine into the competency of waters of atmospheric origin to effect the concentration of these metals.

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[1] Cfr. for instance the repeated finds of galena and other minerals in deep borings on the Gulf Coast. See A. C. Veatch and G. D. Harris, *Bull.* 7, Louisiana Geol. Survey, 1908, p. 25; also, G. D. Harris, *Bull.* 429, U. S. Geol. Survey, 1910, p. 45.

Gilbert Van Ingen has pointed out the frequent occurrence of grains of galena and sphalerite in fossils where perhaps decaying organic matter might have brought about precipitation, *Bull.* Geol. Soc. Am., vol. 26, 1915, p. 85.

In the first place, the mineral association indicates a shallow deposition at temperatures and pressures not very different from those prevailing at the surface. The deposits contain no substances carried by thermal waters of volcanic origin, and no primary silicate minerals. The marcasite suggests strongly deposition near the surface. Barite would be easily concentrated from the limestones. Fluorite is rare in these deposits.

Regarding nickel and cobalt, it has already been pointed out that minerals of these metals are not uncommon in sedimentary strata, as is shown, for example, by their occurrence with the marine siderites and limonitic oölites, or by the occasional discovery of millerite in limestone. This granted, it remains to account for the two principal metals, lead and zinc. The majority of geologists who have studied these deposits believe that the lead and zinc originally were contained as silicates or sulphides in the older crystalline rocks from which the limestones and other sedimentary rocks were derived. A number of analyses have been made particularly by J. D. Robertson (p. 10), which indicated the presence of copper, lead and zinc. Still more convincing is the analysis by George Steiger of a composite sample of 329 igneous rocks which have been analyzed in the laboratory of the U. S. Geological Survey. This gave in per cent. 0.00513 zinc, 0.00075 lead, 0.00932 copper, 0.00515 nickel and 0.00048 arsenic.[1] Another series of analyses of igneous rocks from England by A. M. Finlayson[2] gave an average of 0.0032 per cent. lead and 0.0016 per cent. zinc.

Sedimentary rocks contain apparently less of metals than igneous rocks. According to J. B. Weems and J. D. Robertson the Cambrian and Ordovician limestones of Missouri average in per cent. 0.00425 zinc, 0.00096 lead, and 0.00126 copper.[3] Considered in conjunction with the composite analyses of silts from the Mississippi River delta (p. 252) these figures in part support the opinion referred to regarding decreasing metal content in successive sedimentations.

No analyses of shales are included among those given above. They can not be safely excluded, however, and it is probable that they will average higher in metal content than the limestones

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[1] E. C. Siebenthal, *Bull.* 606, U. S. Geol. Survey, 1915, p. 67.

[2] Geol. Soc. London, Quart. Jour., vol. 66, 1916, p. 301.

[3] E. C. Siebenthal, *op. cit.*, pp. 79.

as indeed suggested by partial analyses quoted by E. R. Buckley[1] and G. H. Cox.[2]

It is assumed by Siebenthal that the metal contents of the igneous rocks is gradually dissipated in successive sedimentations. This may be offset, however, by the fact that limestones are far more easily leached by waters than the crystalline rocks.

In the publication referred to Siebenthal has compiled all available analyses of foreign and domestic waters and has shown that zinc particularly, but also copper and lead, is contained in many samples of the deeper circulation of meteoric waters. Out of 392 waters from Kentucky analyzed by A. M. Peters 89 contained zinc; of these waters 36 also contained H_2S or Na_2S. Most of these waters were obtained from Silurian or Ordovician formations. Similar, though less extensive data are shown from Missouri waters. The zinc is carried by sulphureted salt waters and by alkaline-earthy carbonate waters, the latter usually containing H_2S or CO_2 or both. That acid waters derived from pyritic shales also contain zinc, copper, lead and nickel is shown by the analyses given on pages 55 and 58, and these certainly demonstrate that the metals may be extracted from sedimentary silt deposits. Siebenthal finally found that reservoir deposits from fifteen deep wells of alkaline or saline type in Missouri, Kansas and Oklahoma contained much iron sulphide as well as zinc, lead and copper, all of them probably also present as sulphides. The dried deposits contained a maximum of 0.6 per cent. zinc, 0.2 per cent. lead and 0.1 per cent. copper. Zinc was present in thirteen samples, lead in eleven and copper in nine. The waters themselves commonly yielded a trace of zinc, the greatest amount found being 0.6 part per million.

According to these investigations[3] which represent the most detailed evidence offered by those who advocate an origin from meteoric waters of the deposits under discussion, the zinc and lead existed as finely disseminated sulphides in the older Paleozoic limestone. Waters containing carbon dioxide decomposed the sulphides with the formation of bicarbonates and hydrogen sulphide. In the presence of carbon dioxide, H_2S is not an effective precipitating agent, but when the moving solutions become

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[1] Missouri Bur. Geology and Mines, vol. 9, 1909, p. 221.

[2] *Econ. Geol.*, vol. 6, 1916, p. 587.

[3] E. C. Siebenthal, *op. cit.*, pp. 42–66.

stagnant in places suitable for deposition, CO_2 would escape and the remaining H_2S precipitated the metals as sulphides.

The chlorides of lead and zinc are far more soluble than the bicarbonates and strong brines of sodium chloride are undoubtedly effective in the transportation of the metals. R. C. Wells found that weak salt solutions decomposed but little zinc sulphide.[1] Stronger solutions might be more active. The theory explained is then based on the leaching of lead and zinc occurring as minutely disseminated sulphides in limestone and shale.

Moresnet.[2]—The Moresnet district in Belgium, Luxembourg, and Prussia, is situated in a region of folded Devonian and Carboniferous limestones and slates cut by several large faults and covered unconformably by Cretaceous beds. In the main the ore follows these dislocations, in part as filled veins, in part as large replacement deposits in limestone at the slate contacts or at the intersection of faults. Dolomitization of the limestone is often mentioned. The ore contains zinc blende, galena, iron sulphides, and calcite, and the galena and zinc blende are often intimately intergrown. Nickel is occasionally present. Masses of calamine appeared near the surface and extended to depths of 160 feet; in some cases, notably at Vieille Montagne, they were of enormous size and reached a depth of 330 feet; the sulphides appeared at depths of 170 to 330 feet, much below the water level.[3] Galena is in general the oldest, pyrite or marcasite the youngest of the minerals; concentric intergrowths, of wurtzite, zinc blende, and galena (schalenblende) are not uncommon.

The quantity of ore is said to diminish in depth, and large amounts of water are found. A considerable part of the world's production of zinc has been obtained from these deposits.

Silesia.[4]—Silesia, a province of Prussia, remains one of the world's most important zinc-producing regions. The ore occurs in Triassic sandstone and limestone, which lie in flat syn-

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[1] *Bull.* 606, U. S. Geol. Survey, 1915, p. 58.

[2] Ch. Timmerhans, Les gîtes métallifères de la région de Moresnet, Liège, 1905, p. 28.

F. Klockmann, Die Erzlagerstätten der Gegend von Aachen, Berlin, 1910. See also text-books of Stelzner and Bergeat and R. Beck.

[3] The oxidation of this deposit may be of pre-Cretaceous age.

[4] G. Gürich, Zur Genesis der oberschlesischen Erzlagerstätten, *Zeitschr. prakt. Geol.*, 1903. pp. 202–205.

A. Sachs, Die Bildung der schlesischen Erzlagerstätten, *Centralblatt f. Min.*, 1904, pp. 40–49; *Zeitschr.* Deutsch. geol. Gesell., vol. 56, 1904, pp. 269–272 See also text-books of Stelzner and Bergeat and R. Beck.

clines (Fig. 135). In the lower part of the "Muschelkalk" extensive dolomitization has taken place, mainly along fissures, and the dolomite is underlain by an impermeable "Sohlenstein" or clay rock. The ground-water circulation, dolomitization, and mineralization are all apparently closely connected. The replacement ore occurs along two horizons—the lower in a bed of dolomite several meters thick, carrying galena, zinc blende, and abundant marcasite, and the upper in a bed of smithsonite (zinc carbonate) of considerable thickness. The smithsonite and calamine are considered products of oxidation. The zinc blende is in part intergrown with wurtzite. The galena contains a little silver (0.02 to 0.03 per cent.); manganese as psilomelane is sometimes present. The ores are said to contain on the average about 17 per cent. zinc and 5 per cent. lead. The marcasite contains a little arsenic and a trace of nickel. The succession is marcasite (oldest), zinc blende and galena.[1]

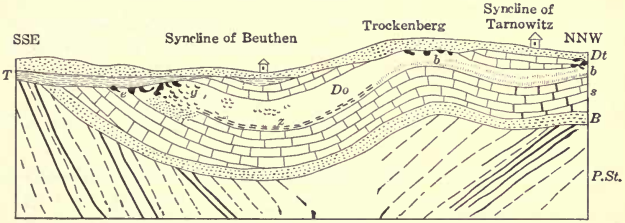

FIG. 135.—Section through the synclines of Tarnowitz and Beuthen, Silesia. *P. St.*, Carboniferous; *B*, Triassic sandstone; *s*, Triassic limestone; *Do*, dolomite; *b*, galena deposits; *z*, zinc blende deposits; *o*, oxidized zinc ores; *e*, limonite; *T*, Tertiary beds; *Dt*, Quaternary beds. *After Gürich.*

There has been much discussion concerning the origin of these ores. Beyschlag and Michael[2] have shown that some ore-bearing fissures descend into the Carboniferous and hence believe that ascending waters did the work; others, like Sachs, believe that the ores resulted from descending waters and that organic matter caused the precipitation.

Alpine Trias.—The Alpine Trias in Austria contains a number of deposits of this type. At Bleiberg, in Carinthia, the ores

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[1] E. Schulz, Geol. Rundschau, vol. 4, 1913, pp. 126–136.

[2] Beyschlag, *Zeitschr. prakt. Geol.*, 1902, p. 143.

Michael, *Zeitschr.* Deutsch. geol. Gesell., vol. 56, 1904, Protocol, pp. 127–139.

occupy filled flats and gash veins; they consist of light-colored zinc blende and marcasite, with calcite and barite gangue, and a little anhydrite and fluorite, but no quartz. No silver, antimony, copper, or arsenic is present.

At Raibl, made famous by Posepny's investigations,[1] the ores form fillings and replacements along three dislocations. The minerals are sphalerite, occasionally with wurtzite, and galena, with a little marcasite and chalcopyrite, and their deposition was accompanied by extensive dolomitization. Posepny describes stalactites of galena, pyrite, and zinc blende, but such occurrences are exceptional.

Other European Localities.—The great deposits of Santander, Spain, are contained in Carboniferous limestone and are said to be replacements connected with dislocations. The light-yellow zinc blende from these deposits is famous. Some cinnabar is present.

At Monteponi,[2] Sardinia, large "stocks" of galena with zinc blende and pyrite are contained in Paleozoic limestones. There is much dolomitization, and a little quartz and barite also occur. Cinnabar is reported and the ores contain silver in part. Igneous rocks are represented only by a diabase.

The genetic relations of both of these deposits are as yet uncertain.

The Lead-Zinc Ores of the Mississippi Valley.—One of the most remarkable metallogenetic provinces characterized by lead and zinc ores extends over the valley of the Mississippi in the generally flat-lying limestones of the Paleozoic, ranging from the Ordovician to the lower Carboniferous (Mississippian) inclusive. These ores are found in Arkansas, Missouri, Oklahoma, Kansas, Illinois, Wisconsin, and Iowa and reach eastward as far as western Virginia and Tennessee. The ores are mined on a large scale in comparatively few regions. Small deposits of lead and zinc are widely spread and are even found in Pennsylvania, New York and Ontario. Igneous rocks are absent. There are, however, a few small deposits in southern Arkansas, Kentucky, and southern

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[1] F. Posepny, *Jahrb.* K. k. geol. Reichsanstalt, vol. 23, 1873, pp. 315–420.

The genesis of ore deposits, *Trans.* Am. Inst. Min. Eng., vol. 23, 1894, pp. 197–369.

An elaborate series of illustrations of this deposit was recently published by the Department of Agriculture of Austria.

[2] R. Beck, Lehre von den Erzlagerstätten, vol. 2, 1909, p. 257.

Illinois in which gold, silver, antimony, or fluorite is present and which appear to be genetically related to local intrusions of igneous rocks. The main characteristics of the predominating type are sufficiently described in the introduction to this chapter. In details they differ considerably.

In point of production[1] the deposits in Missouri easily predominate. The zinc-mining industry centers in the southwestern part of that State, about Joplin, and in 1917 yielded 132,730 short tons of spelter, of a value of about $27,000,000, making about 20 per cent. of the production of the United States. The lead mining in the southeastern part of the State in the same year produced 204,545 short tons of lead, to which should be added 29,611 tons from the Joplin region, making a total value of $40,000,000. This is 37 per cent. of the lead production of the United States. The ore mined is generally referred to as "dirt;" the concentrates are spoken of as "ore." The total quantity of crude ore raised annually in Missouri is now about 19,000,000 tons, consequently it is of low grade. Practically all of it is treated in concentrating works, to yield high-grade material suitable for the reduction plants.

Southwestern Missouri.[2]—The Joplin region includes, outside of Missouri, adjacent parts of Kansas and Oklahoma. The principal camps are at Aurora, Granby, Webb City, Alba, Neck, Joplin, Galena, Badger, Quapaw, and Miami. In the early days lead was the only metal won, but since 1870 zinc ores have been mined and now predominate entirely. From the districts in Missouri, near Joplin, the quantity of lead recovered is about one-fourth as much as zinc. The yield of lead and zinc concentrates from the crude ore averages, according to Siebenthal, about 3.7 per cent. corresponding to 1.9 per cent. zinc. The

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[1] C. E. Siebenthal and J. P. Dunlop, in Mines Report of Missouri, *Mineral Resources*, U. S. Geol. Survey. Annual publication.

[2] W. P. Jenney, *Trans.* Am. Inst. Min. Eng., vol. 22, 1894.

A. Winslow, Missouri Geol. Survey, vols. 6 and 7, 1895.

E. Haworth, Relations between the Ozark uplift and ore deposits, *Bull.*, Geol. Soc. Am., vol. 11, 1900, pp. 231–240.

H. F. Bain (with C. R. Van Hise), Preliminary report on the lead and zinc deposits of the Ozark region, *Twenty-second Ann. Rept.* U. S. Geol. Survey, pt. 2, 1901.

W. S. T. Smith and C. E. Siebenthal, *U. S. Geol. Atlas*, Folio 148 (Joplin).

E. R. Buckley and H. A. Buehler, The geology of the Granby area, Missouri Bur. Geol. and Mines, vol. 4, 1909.

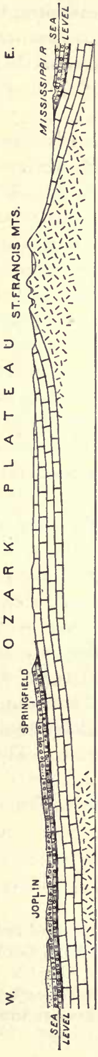

Fig. 136.—General east-west section through Joplin, Missouri, and the St. Francis Mountains, showing the flat-topped dome of Cambrian and Ordovician limestones and pre-Cambrian crystalline rocks overlain by thin Devonian shale and Carboniferous sediments. *After W. S. T. Smith and C. E. Siebenthal, U. S. Geol. Survey.*

concentrates, averaging 58 per cent. zinc, are smelted in Kansas and Oklahoma.

The districts are situated on the flanks of the Ozark uplift (Fig. 136). The ore occurs in the Boone formation and in rocks of Kinderhook age, both belonging to the lower Carboniferous (Mississippian). The beds form a very flat anticline pitching gradually northwest and are displaced slightly by the Seneca fault in Missouri and Oklahoma as well as by the Miami fault in Oklahoma and Kansas. All three structural features appear to be of importance in the ore deposition. The Boone formation contains much light-colored chert, especially in the Grand Falls chert member, which contains the so-called "sheet ground" deposits. The surface of the Boone contained numerous sink-holes and caves, perhaps also drainage channels, and over this "Karst" topography were deposited the sandstones and shales, in part carbonaceous, of the Coal Measures (Pennsylvanian); there is, then, an unconformity by erosion. Post-Carboniferous erosion has now removed much of these rocks, but near Joplin the Pennsylvanian shale still remains in many of the old depressions (Fig. 137).

The succession of the ore minerals is given by Smith and Siebenthal as follows: Dolomite (oldest), chalcopyrite, galena, sphalerite, galena, chalcopyrite, marcasite, pyrite, calcite, barite, and marcasite, the whole series of course being seldom found in one locality. All the minerals are frequently well crystallized. There is general agreement among the investigators that the mineralization began by dolomitization, and Bain sees in it a result of the more rapid diffusion of the magnesia in the ore-forming solutions than of the zinc. The sphalerite occurs as crystals and grains in the secondary chert which forms the gangue of the ore, the primary chert containing no metasomatic sphalerite (Fig. 138). This secondary chert largely made up of cryptocrystalline or microcrystalline quartz, contains much organic matter with minutely disseminated sulphides,[1] and is darker than the primary chert which antedated ore deposition. The dark chert is probably in part a replacement of limestone, in part, where cementing breccias, a silicified mud.

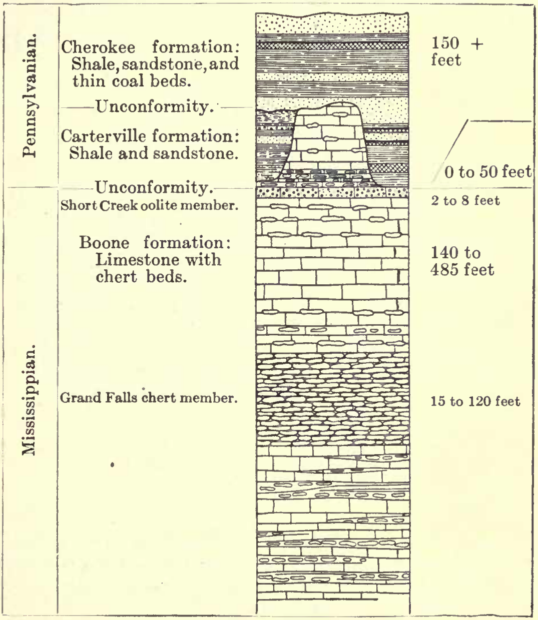

FIG. 137.—Generalized section for the Joplin district, Missouri. *After Smith and Siebenthal, U. S. Geol. Survey.*

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[1] Cox, Dean and Gottschalb, Studies on the origin of Missouri cherts and zinc ores, *Bull.*, School of Mines and Met., Nov., 1916.

A composite sample of the zinc concentrate representing 3,800 lots has the composition given below.[1]

ANALYSIS OF CONCENTRATED ZINC BLENDE FROM THE JOPLIN REGION

Zinc	58.260	Sulphur	30.720
Cadmium	0.304	Calcium carbonate	1.880
Lead	0.700	Magnesium carbonate	0.850
Iron	2.230	Barium sulphate	0.820
Manganese	0.010	Silica	3.950
Copper	0.049		
			99.773

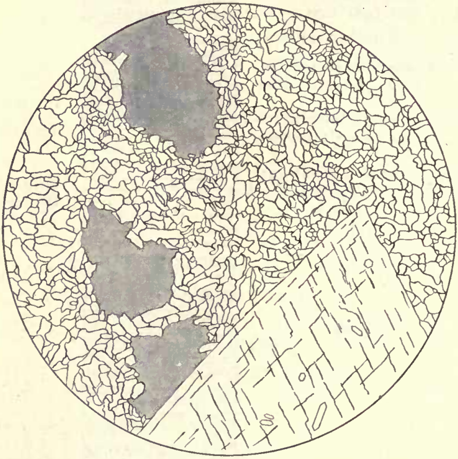

FIG. 138.—Thin section of "black chert" showing matrix of fine-grained quartz with grains of zinc blende (shaded) and crystals of dolomite. Note quartz crystals developing in dolomite. Magnified 53 diameters. *After Smith and Siebenthal, U. S. Geol. Survey.*

The galena contains only a trace of silver.

The ores are found as irregular deposits in the "broken ground" near the surface and as a flat "blanket deposit" or "sheet ground" in a chert member of the Boone formation at depths of 150 to 300 feet. Below this horizon there are, as yet, unworked

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[1] W. G. Waring, The zinc ores of the Joplin district, *Trans.*, Am. Inst. Min. Eng., vol. 57. 1918, pp. 657–670. Waring has also found thallium, indium, gallium and germanium in the flue dust and in the zinc metal.

deposits of disseminated ore of doubtful value. The ore occurs mainly as fillings of cavities, the fillings of distinct veins or crevices being subordinate. The ore minerals with secondary chert fill spaces of brecciation or solution cavities along the stratification, perhaps also spaces of discission in limestone caused by stretching and adjustments.

In the "broken ground," which extends for 100 or 150 feet below the surface, the ores occur in clayey chert breccias in old sink holes filled with Pennsylvanian sediments, or along the outside of such sink holes, forming "circles" where the slipping and settling provided open ground (Fig. 139). In these occurrences the galena predominates, partly because of solution and oxidation of zinc blende, and partly because the deposition of galena prevailed at these upper levels, below the shale. Large masses of galena are found here, in contrast to the conditions in the sheet ground.

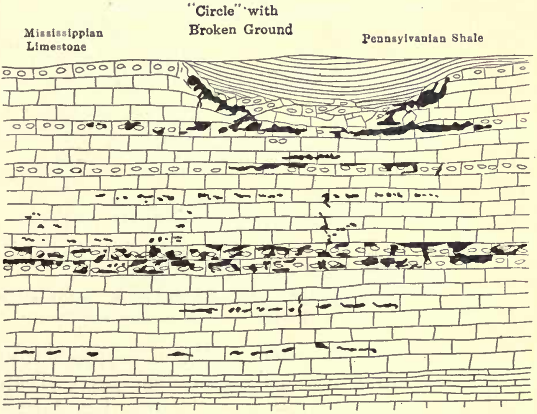

FIG. 139.—Diagram of zinc-lead deposits at Joplin showing "broken ground" around "circle" near surface and "sheet ground" deposit in Grand Falls chert member below. Black areas represent ore. Scale 100 feet to one inch.

Both at Joplin and at Aurora (Fig. 140), as well as in camps in Oklahoma, the "runs" are also a characteristic form of the upper deposits; these sometimes extend for 1 or 2 miles, following the same horizon at depths less than 150 feet—usually much less. At Granby the width of the run is rarely more than 50 to 150 feet. Each run has usually several "openings" (brecciated ground filled with ore), each opening being rarely more than 5 or 6 feet thick. These runs appear to be solution cavities controlled by joints in the rocks.

While in places the brecciation and mineralization continue down to the blanket deposits of the sheet ground, the latter extends in the main independently of the old pre-Pennsylvanian surface. In this sheet ground, which is from 6 to 15 feet thick, the galena and zinc blende occur in dark chert, filling brecciated old chert, and in elongated, narrow solution cavities due to dissolved streaks of limestone in the prevailing chert. The sheet ground is the most important source of ore (Fig. 141).

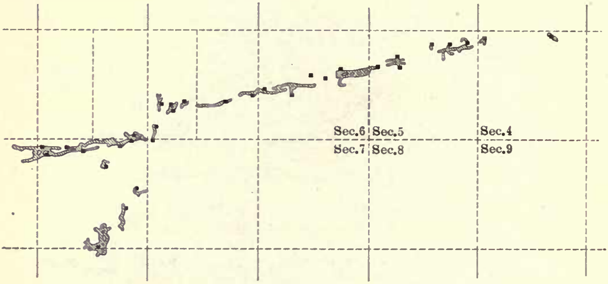

FIG. 140.—Plan showing shafts and workings along run of galena ore, north of Aurora, Missouri. *After Arthur Winslow.*

The newly discovered deposits in Oklahoma[1] at Miami and Quapaw have assumed great importance. The ore bodies form "runs" which in the main extend N.E. approximately parallel to the Miami fault. The ore which is richer than at Joplin occurs at a depth of 200 feet or less in Mississippian rocks underneath a rather thick cover of Pennsylvanian shales. The mine water appears to form part of an artesian circulation and contains much H_2S. Bitumen is in part so abundant as to become an objectionable constituent. The whole occurrence appears to support the theory of origin by ascending waters.

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[1] E. C. Siebenthal, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 187–228.

Maps by H. A. Buehler in "War minerals of the Joplin district," Am. Inst. Min. Eng., Joplin-Miami meeting, Oct., 1917.

Arkansas.—In northern Arkansas,[1] a short distance southeast from the Joplin region, the zinc blende, generally without galena, occurs in fissures or crevices, in fault breccias, and in solution breccias, accompanied by secondary chert or dolomite, sometimes also by crystallized quartz; the ores are found in Ordovician limestone and also in the Boone (Mississippian) formation.

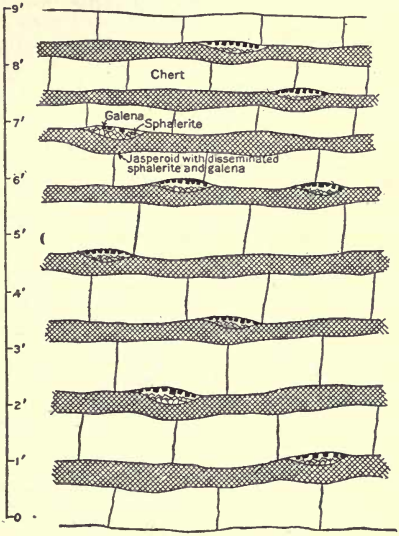

FIG. 141.—Sketch illustrating the occurrence of galena and sphalerite in cavities in the sheet ground, Joplin, Mo. *After C. E. Siebenthal, U. S. Geol. Survey.*

Upper Mississippi Valley.[2]—The districts of the upper valley

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[1] G. I. Adams, *Prof. Paper* 24, U. S. Geol. Survey, 1904.
J. C. Branner, Arkansas Geol. Survey, vol. 5, 1900.

[2] J. D. Whitney, Geology of Wisconsin, vol. 1, 1862.
T. C. Chamberlin, Geology of Wisconsin, vol. 4, 1882.
W. P. Jenney, *Trans.*, Am. Inst. Min. Eng., vol. 22, 1894, pp. 208–209
C. R. Van Hise, Some principles controlling the deposition of ores, *Trans.*, Am. Inst. Min. Eng., vol. 30, 1901.
H. F. Bain, *Bull.* 294, U. S. Geol. Survey, 1906.
U. S. Grant, *Bull.* 14, Wisconsin Geol. Survey, 1906.
G. H. Cox, *Econ. Geol.*, vol. 6, 1911, pp. 427–448; 582–603.
H. C. George, *Bull.* 132, Am. Inst. Min. Eng., 1917, pp. 2045–2074.

lie in Wisconsin, Iowa, and Illinois. The most important districts are in Wisconsin and yielded in 1917 about 4,100 short tons of lead and 59,700 tons of zinc, with a total value of about $13,000,000. The ore deposits are found in Ordovician strata of almost horizontal position. The following formations are recognized:

	Feet
Cincinnati or Maquoketa shale	160
Galena dolomite	240
Platteville limestone (Trenton)	55
St. Peter sandstone	100
Lower magnesian limestone	350

Below the magnesian limestone is 700 feet of the Cambrian Potsdam sandstone. The so-called "oil rock," a thin bed of bituminous shale, is found at the base of the Galena formation or at the top of the Trenton. The rocks dip gently southwest and are flexed into very shallow troughs.

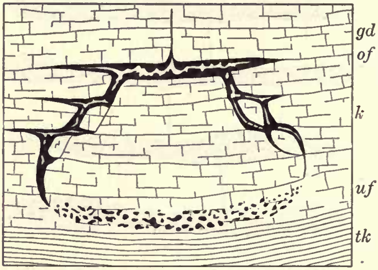

FIG. 142.—Section showing occurrence of lead and zinc in vertical crevices, flats, and pitches; also of disseminated ores in the rocks. *gd*, Galena dolomite; *tk*, Trenton limestone; *of*, upper flat; *uf*, lower flat; *k*, connecting flats, pitches, and verticals. *After T. C. Chamberlin.*

The ores are confined to the Galena limestone and the upper part of the Platteville limestone; the minerals consist of marcasite, sphalerite, and galena, deposited in the order noted. The gangue is crystallized calcite, rarely barite. Cadmium is absent, but a trace of silver is found. The abundance of marcasite causes metallurgical difficulties and necessitates treatment of the concentrates in electrostatic or magnetic separators, in the latter case preceded by partial roasting. The ores occur as fillings of open spaces, vertical crevices, or "gash veins" connected with "pitches" or "flats," all probably due to solution along joint planes (Fig. 142). Stalactites of sulphides are sometimes found indicating that the spaces were not always filled by solutions. In part there are also flat bodies of disseminated ores. The galena predominates near the surface, probably largely because the zinc blende has been dissolved as sulphate and transformed to silicate or carbonate in the lower levels (p. 455). In depth zinc blende with a little galena is the principal ore. Mining operations extend to a depth of at most 200 feet. The distribution of the oil shale, according to Bain, seems to coincide with the extent of the deposits. Cox, however, holds that the metal was derived from the overlying Maquoketa shale and carried down to be concentrated in the Galena limestone.

Virginia and Tennessee.[1]—In western Virginia, and near Knoxville, Tennessee, lead and zinc ores occur in the Cambro-Ordovician (Shenandoah) limestones, mostly where the rocks are faulted or brecciated or where they carry much organic matter. The gangue consists of calcite, dolomite, and rarely barite. There is little quartz or pyrite and no definite order of crystallization.

Southeastern Missouri.[2]—In eastern Missouri not far from the Mississippi River and south of St. Louis lead mining has been carried on more or less extensively since the early part of the eighteenth century, but in the last ten years the industry has assumed very large proportions. In 1917 the yield of lead from this region was 204,545 short tons, worth more than $35,000,000. The crude ores, which yield on the average 5.5 per cent. of lead concentrates, are treated at the rate of 20,000 tons per day in local concentrating works and a part of the galena

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[1] T. L. Watson, Lead and zinc deposits of Virginia, Geol. Survey Virginia, vol. 57, 1905.

Frank L. Nason, Characteristics of zinc deposits in North America, *Trans.* Am. Inst. Min. Eng., vol. 57, 1918, pp. 830–855.

H. A. Coy and H. B. Henegar, Mining methods of the American Zinc Co. of Tenn., *idem*, vol. 58, 1918, pp. 36–47.

[2] A. Winslow, Missouri Geol. Survey, vols. 6 and 7, 1894.

A. Winslow, *Bull.* 132, U. S. Geol. Survey, 1896.

C. R. Keyes, Missouri Geol. Survey, vol. 9, 1896.

E. R. Buckley, Geology of the disseminated lead deposits, Missouri Bur. Geol. and Mines, vol. 9, pts. 1 and 2, 1909.

A. P. Watt, Concentration practice in southeastern Missouri, *Trans.*, Am. Inst. Min. Eng., vol. 57, 1918, pp. 322–419 is smelted in the district. Practically no zinc is contained in the ore.

The geological position of the deposits is in the Cambrian and therefore lower than those of the other Mississippi Valley ores. On an irregular surface of pre-Cambrian granite and porphyry rests the basal La Motte sandstone, about 200 feet thick (Fig. 143). Above this lies the arenaceous dolomite of the Bonneterre formation, often chloritic, with beds of shale having in all a thickness of 300 to 400 feet. Covering the Bonneterre are the Davis, Derby, Doe Run, and Potosi formations, which are mainly dolomites and shales and all of which belong to the Upper Cambrian.

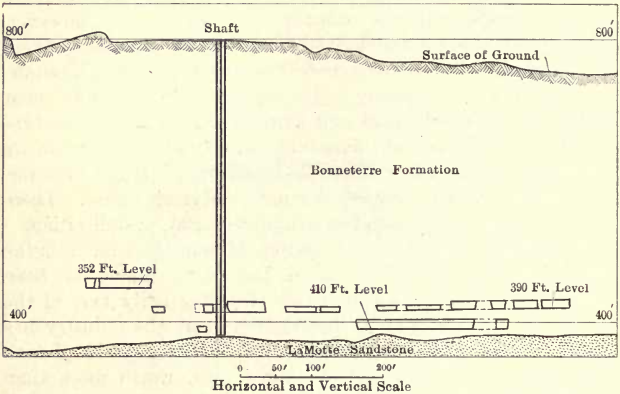

FIG. 143.—Vertical section showing workings in mine No. 4, Federal Lead Company, southeastern Missouri. Horizontal bodies of disseminated ore, following bedding of shaly dolomite of Bonneterre formation. *After E. R. Buckley.*

The principal ore horizon is in the lower part of the Bonneterre dolomite, though some galena occurs throughout that formation. A second, less important ore horizon is in the Potosi dolomite, where the galena is accompanied by barite. The strata are horizontal or have very gentle dip.

The ore minerals are mainly galena accompanied by calcite, a little pyrite, and sometimes chalcopyrite. In places—for instance, at Mine La Motte and Fredericktown—the ores contain nickel and cobalt, as linnæite $(Co,Ni)_3S_4$; some of the ores have been worked for these metals. Watt quotes a representative analysis of the crude ore of the southeastern district.

ANALYSIS OF DISSEMINATED ORE FROM SOUTHEASTERN MISSOURI

Pb	4.32	Al_2O_3	1.16
Cu	0.03	CaO	30.80
Zn	0.50	MgO	17.96
S	0.97	CO_2	32.79
SiO_2	4.83		
Fe_2O_3	6.64		100.00

Silver 0.12 oz. per ton. Trace Ni, CO, Mn.

Watt states that the silver follows the zinc. Concentrates of zinc blende contain up to 10 ounces of silver per ton.

The ores are often called disseminated, for the galena usually occurs as grain or crystals disseminated in the greenish-gray dolomite (Fig. 144); sometimes these crystals are several centimeters in diameter. According to Buckley the ores of the lower part of the Bonneterre occur as follows:

1. As horizontal sheets along bedding planes, generally along the upper side of thin shale beds.
2. Disseminated in dolomite.
3. Filling or lining joints.
4. In cavities or vugs.

The galena is persistently associated with dark dolomite and black shale.

The ores are mined from vertical shafts, 100 to 550 feet deep. The ore does not extend in all directions like a coal bed, but the flat shoots or "runs" follow rather persistently one or two directions, undoubtedly controlled by joints and small faults. Some of these runs have been followed for miles and may be several hundred feet wide; some of the mine workings in the Bonneterre district are 100 feet high.

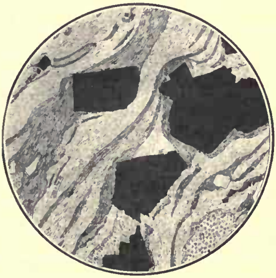

FIG. 144.—Crystals of galena developing in shaly dolomite. Black, galena; shaded and stippled, shaly dolomite; white, quartz. Magnified about 10 diameters. *After E. R. Buckley.*

Genesis of the Mississippi Valley Deposits.—An unusually extensive literature, full of controversy and divergent views, covers the question of the genesis of these ores. A majority of the authors agree that the source of the ores was in the Paleozoic sedimentary beds, also that the deposition was effected by atmospheric waters, and finally that the metals were in solution mainly as sulphates. Summaries of the various views are found in the text-books of Kemp and Ries and in the reports of A. Winslow and E. R. Buckley. At present there are two strongly contrasting opinions regarding the Missouri deposits. The descensionists are represented by Whitney, Chamberlin, Blake, Robertson, Winslow, Buckley, and Buehler and the ascensionists by Jenney, Nason, Van Hise, Bain, W. S. T. Smith, and Siebenthal. From the latter we may separate Jenney and Nason, who see in the ore deposits the result of fissuring extending into the underlying pre-Cambrian rocks, through which thermal waters ascended. Buckley and Buehler hold that the source of the lead and zinc was in the Pennsylvanian sediments, which, however, contain no important deposits and only in places small amounts of galena and zinc blende. The finely distributed sulphides were dissolved as sulphates and carried downward in acid solutions which finally mingled with neutral or alkaline solutions from the unoxidized parts of Pennsylvanian sediments. These mingled waters deposited galena and zinc blende in the sink holes and drainage channels of the underlying Mississippian limestone and chert. Unlike Winslow, Buckley and Buehler do not believe that the deposition was effected at the time of deposition of the Pennsylvanian shale, but later, after the erosion of a part of those beds. The obvious difficulty in their theory appears to be that it requires the waters to have descended through an impervious shale cover.

A similar theory is advocated by Buckley for the lead deposits of southeastern Missouri. The pre-Cambrian rocks are held to be the original source of the metals. The water flowing into the Cambrian sea contained lead, which was deposited with the Bonneterre dolomite as small particles. After the subsequent formations were laid down the concentration of lead by surface waters began. Finally the Pennsylvanian shales were laid down over this area and from them the greatest amounts of metals were derived. The solutions were thus in the main descending, though in part they may have ascended in artesian circulation through the La Motte sandstone. Buckley states, indeed, that even at present there are strong indications of artesian conditions in the mines. On the whole the ore deposition is post-Pennsylvanian. Here again the impermeable character of the Pennsylvanian may be advanced as an argument against Buckley's view, as well as the improbability of a strong descending flow through the great thickness of Cambrian, Ordovician, and Mississippian beds. A satisfactory explanation of the southeastern Missouri deposits is as yet lacking.

On the other hand, Van Hise, Bain, Tangier Smith, and Siebenthal, who have studied the Joplin district, believe that the source of the ores is in the various formations below the Pennsylvanian, particularly in the Cambro-Ordovician, and that atmospheric waters penetrating these rocks were carried up against the impervious beds of the Pennsylvanian and here deposited in the pre-Pennsylvanian breccias and sink-holes. Smith and Siebenthal hold that the ores were formed much later than the Pennsylvanian, after the Ozark uplift (Fig. 136) had established an artesian circulation. The surface waters entered the older Paleozoic outcrops to the south and east of the Joplin region. After following these beds they passed upward through the jointed and brecciated Mississippian limestone until they reached the vicinity of the impermeable shales. There is, indeed, in the deep wells of Joplin good evidence of the existence of artesian pressure. Siebenthal's recent contribution in which very strong arguments are advanced in favor of the artesian theory has already been reviewed in the general part of this chapter.

For the deposits of Wisconsin and Illinois, Van Hise and Bain assume that the metals were minutely disseminated as sulphides through the Galena dolomite and concentrated, probably in late Tertiary or post-Tertiary time, by the action of surface waters descending in shallow troughs through the fractured and slightly inclined Galena limestone, and that the reduction was effected by the organic matter of the oil rock. In the Galena limestone the solutions were practically confined between two beds of almost impermeable shale.

According to Cox the metals were derived from the overlying Maquoketa shale, in which he finds some evidence of the presence of sulphides, particularly sphalerite. He points out, with good reason, that the shales are better suited as receptacles for metallic ores than the limestones. The latter are deposited in deeper water, while the shales are shore formations in which the metallic substances derived from adjacent continents would most easily be deposited as detritus or precipitated from solutions.

We find here the same conflict of opinion, as in the case of the Missouri deposits and those of Silesia, between the ascensionists and the descensionists. The problem is not yet solved, but looking beyond these controversies, we cannot deny that in many countries transition types appear which seem to connect these apparently distinct non-igneous deposits with deposits of igneous affiliations.

L. V. Pirsson[1] in 1915 expressed the view that the zinc-lead deposits of the Mississippi valley type might well result from "the quiet upward movement of volatile magmatic material" thus ranging himself with W. P. Jenney and F. L. Nason. It will be incumbent upon the supporters of this theory to controvert the strong arguments offered by Siebenthal in favor of leaching of limestones by ascending saline meteoric waters.

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[1] Origin of certain ore deposits, *Econ. Geol.*, vol. 10, 1915, pp. 180–186.

CHAPTER XXIV

METALLIFEROUS DEPOSITS FORMED NEAR THE SURFACE BY ASCENDING THERMAL WATERS AND IN GENETIC CONNECTION WITH IGNEOUS ROCKS

CHARACTER AND ORIGIN

General Features.—The deposits at the orifices of hot ascending springs have been described in Chapter VII. It has been shown that they consist of opal, chalcedony, quartz, calcite, aragonite, barite, and fluorite, with a number of other gangue minerals, and that they also contain in places metallic gold and certain sulphides, such as cinnabar, stibnite, and pyrite, but not the other common ore minerals such as chalcopyrite, galena, zinc blende, and arsenopyrite. The ore deposits described in the present chapter present some striking analogies to those products of the hot springs.

In regions of comparatively recent volcanic activity where the measure of erosion since the eruptions ceased is in hundreds rather than in thousands of feet we find a group of important ore deposits, most commonly in the form of fissure veins. They generally occur in igneous flow rocks[1] and also cut the underlying or adjacent formations. They constitute the source of a large part of the world's production of gold, silver, and quicksilver, and they contain the spectacular bonanzas of the Cordilleran region, of which examples are found at Tuscarora, Virginia City, Goldfield, Cripple Creek, Pachuca, Guanajuato, and many other districts. Following the Tertiary outbursts of effusive rocks, these deposits accompany the "circle of fire"

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[1] We are accustomed to consider as intrusive rocks those which have congealed with granular texture far below the surface. Intrusions are, however, not confined to any particular depth or texture. Intrusive bodies may be found in any series of rocks even near the surface and may then have fine-grained, trachytic or even glassy texture. The distinction between flows and intrusions may in such cases become difficult and, as shown in case of the Tonopah, Waihi and other districts, the relations may have far reaching bearing upon the richness and continuation of the deposits contained in such a series of rocks.

that encompasses the Pacific Ocean. We find them in Japan, in the East Indian Islands, and in New Zealand. They are characteristically developed in that classical mining region of the Old World, in Hungary and Transylvania, where one of the elements—tellurium—which so often accompanies them was first found.

Though most of these ore-deposits are found in the Tertiary flow rocks they are not confined to rocks of this period. There is good reason to believe that veins are developing now in some regions of recent volcanism, and also that similar veins have been formed during pre-Tertiary outbreaks, although erosion has removed most of the older representatives of this type. These deposits have certain well-marked characteristics which are partly of a mechanical, partly of a chemical origin.

Because the fissuring of the rocks took place near the surface, under slight load, open cavities were abundant, and filling, crustification, and comb structure are conspicuous. The walls are likely to be irregular, and the vein matter is often "frozen" to the walls. Splitting, chambering, and brecciation are features of the veins. While metasomatic processes have been active in the surrounding rocks, the ore is usually confined to the open fissures. Short and irregular veins are more frequent than the regularly developed conjugated fractures resulting from strong compressive stress. Divergent systems of fractures or several parallel systems with little apparent relationship are thought to be due to the gravitative settling of volcanic piles.

Banding caused by crustification is common, as illustrated in Figs. 160, 166, 172, 176 and 180. It is much more delicate and frequent than in deposits formed at greater depth and higher temperature.

The occurrence of the ore in "stockworks," or in pipes, or below impervious beds is often observed. In superimposed lava flows of different kinds, some are usually better adapted to the deposition of ore than others and this difference may result in the development of ore-shoots which are approximately horizontal.

Among the metals contained in these deposits gold and silver are by far the most important. Base metals are present, plentifully enough in places, but the mines are rarely worked for these. Large bodies of galena and zinc blende occur in some places, but it is decidedly rare to find important copper deposits. The "pyritic" deposits are not represented; they are confined to the deeper zones or to those of higher temperatures. Arsenic and antimony, bismuth, tellurium, and selenium are common but are rarely of economic importance; quicksilver is present in some deposits and indeed the typical quicksilver deposits belong to this class. Cobalt and nickel, tungsten, and molybdenum are not unknown, but are entirely subordinate. Their home is in the deeper deposits.

The pure gold deposits are relatively scarce. Those carrying silver only are common in certain regions, like Mexico. The usual metals are gold and silver occurring together in varying proportions.

Among the ore minerals native gold should be mentioned first. It contains silver, as a rule, and is of pale yellow color; a proportion sometimes occurring is ounce for ounce when the mineral is of very pale grayish-yellow color (electrum). Deep yellow gold is not unknown, however. The gold is often present in very fine mechanical distribution, being sometimes so closely intergrown with ore minerals and gangue that no colors can be obtained by panning. When derived by oxidation of tellurides it is of dull brown color and is difficult to recognize even in rich specimens. The whole series of tellurides is present. As the gold generally occurs in minute particles rich placers below the croppings of these deposits are rather unusual.

Native silver is ordinarily a product of oxidation. The primary and most abundant silver mineral is argentite; complex silver sulphantimonides and sulpharsenides are also characteristic; it is often difficult to say which are secondary and which are primary. Among them are proustite, pyrargyrite, miargyrite, stephanite, polybasite, tetrahedrite, and more rarely enargite.

Stibnite is plentiful in deposits of certain types. Among the base minerals pyrite is always present, but in small quantity and fine distribution; marcasite, a mineral typical of deposition near the surface, is not unusual; often it is secondary. There are also galena, zinc blende, chalcopyrite, and sometimes alabandite; rarely arsenopyrite; never pyrrhotite or magnetite.

Of gangue minerals quartz is the most abundant, and crystals of it are plentiful but rarely large; an amethyst color is often noticeable. The quartz aggregates are not glassy or milky but usually fine-grained (hornstone) and often chalcedonic, with banding and rapidly changing grain. Chalcedony and opal are usually later than the quartz. Calcite, dolomite, barite, and fluorite are locally the dominant gangue minerals, while siderite is rare. Manganese minerals like rhodochrosite and sometimes rhodonite are typical of certain groups. Kaolin accompanies the veins, sometimes in large amounts, but is probably in most cases a product of secondary changes by descending waters. Sericite and chlorite appear in the altered country rock. Zeolites are present in some deposits, but are certainly of exceptional occurrence.

One of the most widespread and characteristic gangue minerals and the most difficult to explain is adularia (or valencianite).[1] Discovered by Breithaupt in specimens from the Valenciana mine at Guanajuato, this mineral has since been found in numerous other places, mainly in the Cordilleran region, as part of the filling, and as a metasomatic product in the country rock. Among the places where this feldspar plays an important part may be mentioned Silver City (Idaho), Tuscarora, Tonopah, and Rawhide (Nevada), Gold Road (Arizona), Republic (Washington), and Cripple Creek and Creede (Colorado). It does not occur at Goldfield, Nevada, where solutions of acid reaction appear to have deposited the ore. The orthoclase mineral is usually a pure potassium feldspar, although varieties with several per cent. of sodium have been found at Waihi, New Zealand, and in the Gold Spring district, Utah.[2] Sometimes the adularia replaces orthoclase, biotite, and other rock minerals (Fig. 145); it is also found in the form of well-developed crystals of prism and dome intergrown with vein quartz (Fig. 146). The cross-

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[1] A. Breithaupt, Ueber die Felsite und einige neue Specien ihres Geschlechts, *Schweigg. Jour.*, Bd. 60, p. 322, 1830.

W. Lindgren, Orthoclase as a gangue mineral in fissure veins, *Am. Jour. Sci.*, 4th ser., vol. 5, 1898, p. 418.

W. Lindgren, *Twentieth Ann. Rept.* U. S. Geol. Survey, pt. 3, 1900, p. 167.

J. E. Spurr, *Prof. Paper* 42, U. S. Geol. Survey, 1905, p. 86.

W. Lindgren and F. L. Ransome, *Prof. Paper* 54, U. S. Geol. Survey, 1906, p. 187.

A. F. Rogers, Orthoclase-bearing veins from Rawhide, Nevada, *Econ. Geol.*, vol. 6, 1911, p. 790.

F. C. Schrader, Mineral deposits of the Cerbat Range, Black Mountains and Grand Wash Cliffs, Mohave County, Arizona, *Bull.* 397, U. S. Geol. Survey, 1909.

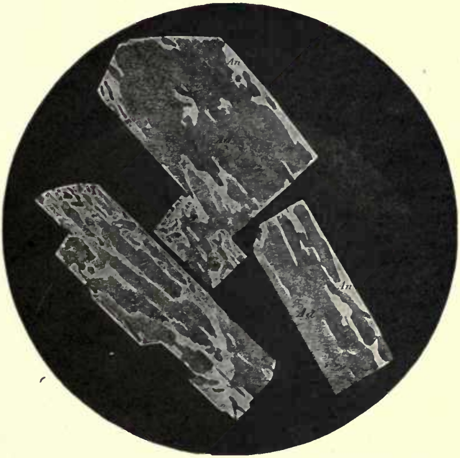
F. C. Schrader, A reconnaissance of the Jarbidge district, Nevada, *Bull.* 497, U. S. Geol. Survey, 1912.

[2] B. S. Butler, U. S. Geol. Survey, oral communication.

sections of the adularia crystals are usually of rhombic shape. The mineral also occurs abundantly in some veins that had originally a calcite gangue, now replaced by an intimate intergrowth of adularia and quartz.

The high-temperature minerals, such as augite, amphibole, olivine, biotite, tourmaline, topaz, garnet, magnetite, ilmenite, and chromite, are conspicuously absent.

Successive Phases of Mineralization.—Veins formed near the surface in volcanic regions are sometimes subject to peculiar changes, which are rarely observed in deposits of more deep-seated origin. An earlier gangue mineral, such as calcite or barite, may be wholly wiped out and replaced by a new gangue of quartz and adularia. This alteration has nothing to do with surface waters though the latter may sometimes produce a similar cellular or lamellar structure; it is plainly caused by a change in the composition of the ascending currents. Indications of this process may be seen even where it has not been carried to completion. In many veins at Cripple Creek deposition began by the growth of slender crystals of celestite from the walls,



.Fig. 145.—Adularia (*Ad*) replacing soda-lime feldspar (*An*) in andesite from Tonopah, Nevada. Magnified 17 diameters. *After J. E. Spurr, U. S. Geol. Survey.*

and these crystals were subsequently replaced by quartz, in which the pseudomorphs are now embedded. In the Trade Dollar vein at Silver City, Idaho, the filling consists of quartz and adularia, but casts of barite or calcite covered with minute crystals of adularia indicate that here also there was a preliminary carbonate or sulphate stage.

In many instances the vein was completely filled by calcite, each grain separated by a slender partition of quartz; at the beginning of the second stage this calcite was dissolved, leaving a skeleton of thin silica walls; secondary quartz and often also adularia were deposited upon these walls, giving them more strength, but the ore remains a delicate aggregate of "hackly" or lamellar quartz, such as is exceedingly characteristic of some mining district. At De Lamar, Idaho, this ore consists only of quartz (Figs. 147 and 148). In the veins at Gold Road, Arizona, and many other veins in the same district, the original gangue material consisted of calcite and fluorite and the "pseudomorphic" ore consists of quartz and large amounts of adularia. Similar ore may be seen in the Mount Baldy district, southern

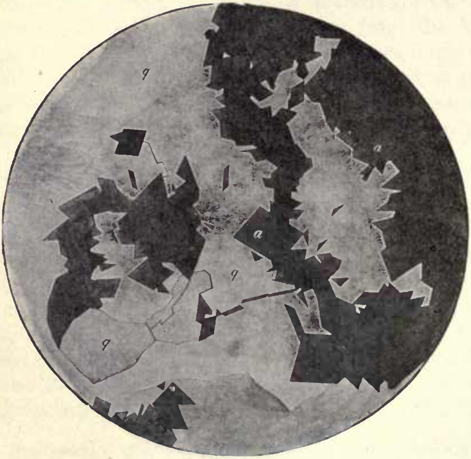


FIG. 146.—Intergrowth of quartz (q) and adularia (a), Fraction vein, Tonopah, Nevada. Magnified 38 diameters. *After J. E. Spurr, U. S. Geol. Survey.*

Utah, at Jarbidge, Nevada (Fig. 149), and at many other places. This important development of adularia, involving transportation of alumina by siliceous solutions, remains without full explanation. The composition of the ore may be similar to that of a pegmatite dike, but the structure is wholly different.

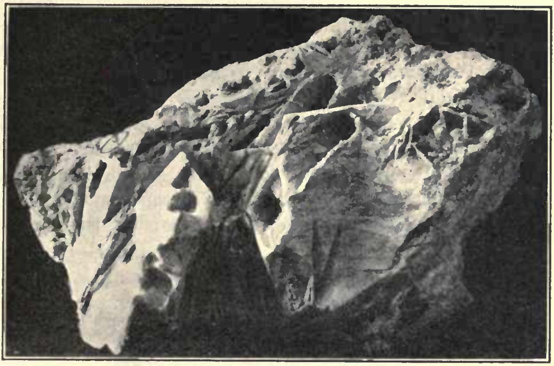


FIG. 147.—Lamellar quartz, replacing calcite gangue, De Lamar, Idaho. One-half natural size.

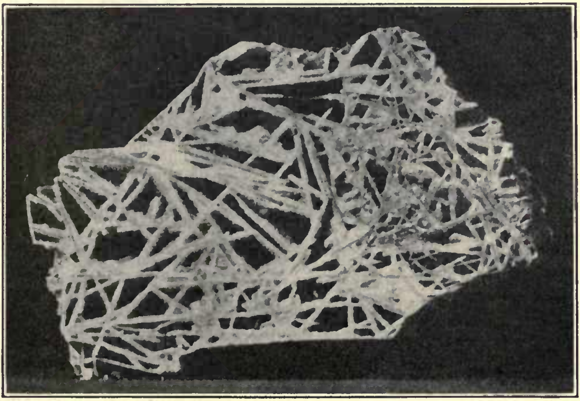


FIG. 148.—Section of lamellar ore, De Lamar, Idaho. Natural size.

There is reason to believe that this "pseudomorphism" is accompanied by a change in the metal content of the vein. At least it seems as if the original filling of barite, calcite, and fluorite carried more silver and as if the silicification and feldspathization was accompanied by a concentration of the gold. Similar processes may be traced in some quartz veins of the Republic district, Washington. Here quartz with some adularia replaces a slender acicular or thin tabular mineral, probably calcite, developed parallel to *c* and *r*, which seems to have been deposited only along the walls of the vein.

Zeolitic Replacement.—Zeolites are foreign to veins of the deeper zones; in the veins formed near the surface they are occasionally found, but they are rare. At a few places zeolites are reported in the altered country rock (Tonopah, the Comstock, and Waihi). At Guanajuato zeolites are found in the filling of the veins, but here they always belong to the latest phases of vein formation. Apophyllite, laumontite, and stilbite are the species reported. Few of these occurrences in the vein filling have been carefully studied. In the Southern Republic mine at Republic, Washington, laumontite, associated with calcite, occurs on a fairly large scale.[1] At this place the ordinary fine-

FIG. 149.—Thin section of lamellar quartz and adularia, pseudomorphic after calcite, Jarbidge, Nevada. Magnified 12 diameters. *After F. C. Schrader, U. S. Geol. Survey.*

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[1] W. Lindgren, *Trans.*, Canadian Min. Inst., vol. 15, 1912, pp. 187–191.

grained banded quartz filling had evidently been dissolved and the laumontite-calcite aggregate was deposited in its place. The ore in this zeolitic zone or shoot contains mainly silver, whereas elsewhere in the mine gold predominates in the quartz gangue. It seems to be worth while to draw attention to this occurrence with a view to ascertaining whether the development of zeolites is not favorable to silver enrichment. Many facts noted in veins of other classes, like those of Kongsberg and Andreasberg, point in this direction.

Primary Ore Shoots, Oxidation, and Sulphide Enrichment.— Rich oxidized ores are often encountered in these deposits at the surface and down to the water-level. Whether the primary ore is greatly enriched in this zone depends really more on the texture and composition of the ore than on its original tenor. In veins of hard fine-grained quartz oxidation often fails to produce an ore of higher grade. There are many districts in which the oxidized ores are little, if any, richer than those below the oxidized zone. Such conditions exist, for instance, at Cripple Creek and at Tonopah.

The largest and richest masses of ore are often found just below the oxidized zone and in general contain sulphides, sulphantimonides, and sulpharsenides. It will suffice to call attention to the great silver bonanzas of Guanajuato and Pachuca; to the Comstock, where in one month silver-gold ores valued at $6,000,000 were extracted; to Tonopah, Nevada, where in three months ore yielding over $3,000,000 was extracted; to the Caledonian mine at Thames, New Zealand, which in one year from a small ore-shoot produced $6,000,000 in gold; to Cripple Creek, Colorado, where in one year from a small area but a considerable number of mines $18,000,000 in gold was produced; to Goldfield, Nevada, where during a recent year over $10,000,000 in gold was recovered from one property, the ore averaging $38.50 per ton, and where, of the phenomenally rich ore shipped in 1907, one carload of 47 tons yielded $600,000 in gold.

At the same time it is well to emphasize the fact that most of these high yields proved ephemeral. The bonanzas were extracted, poorer ore was found in depth, and the mine was abandoned or continued in feeble existence.

These great bonanzas are due in part to primary deposition— in large degree probably to the later, reconcentrating phases of primary deposition; and in part to sulphide concentration by descending waters charged with precious metals from the upper parts of the veins. It should not be overlooked that ore-shoots of primary origin are common enough. Take, for instance, the Cripple Creek gold ores, in which evidence of enrichment is conspicuously lacking; these primary shoots are usually of a markedly irregular form; many of the smaller ore-bodies are likely to follow intersections of fissures. In depth the rich shoots show a tendency to contract, to feather out, or simply to become impoverished. In many cases zinc blende in a quartz gangue appears in depth. On the other hand sulphide enrichment is conspicuously marked, especially in silver veins. In gold-bearing veins the enrichment in gold is likely to be localized in the lower part of the zone of oxidation. The secondary silver minerals are native silver, argentite, ruby silver, stephanite, and polybasite—that is, the same minerals (except the native silver) as occur in primary ore. As a consequence it is often exceedingly difficult to distinguish primary ore and enriched ore, and when in addition to this we recognize that the later effects of primary mineralization may sometimes closely simulate the products of descending surface waters, the difficulties of correct interpretation will be fully realized. The distinction is made easier when the secondary sulphides form a distinct zone immediately below the oxidized part of the lode.

The conditions for the deposition of gold and silver are apparently much more favorable near the surface than at greater depths; on the other hand, deposition took place rapidly and the gold and silver contents of the solutions were doubtless exhausted before they reached the actual surface.

Types of Deposits.—The merging of the various types makes it difficult to establish a rigid classification.

One type, namely, the zeolitic copper deposits in amygdaloid rocks, has been left out of consideration at this time, for it really represents a mineralization of the lava derived from its own body. To gain a general orientation the deposits here described are classified as follows:

1. *Cinnabar Deposits.*—Cinnabar, marcasite, stibnite, hydrocarbons, quartz, opal, calcite.

2. *Stibnite Deposits.*—Stibnite, pyrite, and some other sulphides; also quartz.

3. *Base Metal Deposits.*—Chalcopyrite, galena, zinc blende,

tetrahedrite, in an abundant gangue of quartz, carbonate, or barite. Principal values usually in gold and silver.

4. *Gold Deposits.*—Native gold, alloyed with silver. Subordinate argentite, ruby silver, etc. Quartz.

5. *Argentite-Gold Deposits.*—Argentite, ruby silver, tetrahedrite, etc.; native gold, quartz, calcite.

6. *Argentite Deposits.*—Argentite, ruby silver, tetrahedrite, etc.; quartz or calcite, barite, and fluorite.

7. *Gold Telluride Deposits.*—Gold tellurides, quartz, or quartz and fluorite.

8. *Gold Telluride Deposits with Alunite.*—Gold tellurides, pyrite, alunite, kaolin.

9. *Gold Selenide Deposits.*—Gold selenides, pyrite, quartz, calcite.

Older Representatives of this Class.—The types just enumerated almost wholly represent veins or similar deposits in Tertiary lavas of the Cordilleran or Pacific or Hungarian regions, but a close examination will easily discover examples of similar deposits of a greater geological age. Beck[1] described relatively unimportant deposits in the Mesozoic melaphyres and quartz porphyries at Imsbach, in the German Palatinate, that carry chalcopyrite, galena, and tetrahedrite in a gangue of calcite, barite, and rhodochrosite and are probably ancient representatives of this class. The celebrated veins of Freiberg, or at least three types of them, namely, the "noble quartz formation," the "noble carbonate formation" and especially the "barytic lead formation," should be mentioned in this connection. There seems to be good evidence that these are Carboniferous representatives of veins formed near the surface, although the lavas in which they probably reached the surface are now eroded. Their structure and composition point clearly to a shallow deposition, and were the physiographic conditions in the Erzgebirge fully analyzed the results would probably confirm this view. The "barytic lead formation," for instance, carries barite, fluorite, quartz, and dense quartz as gangue minerals with beautiful crustification, while the ore minerals are argentiferous galena, tetrahedrite, bournonite, and chalcopyrite.

Another occurrence that might well be cited comprises the insignificant veins in the Triassic diabase flows at Bergen Hill, New Jersey, which contain pyrite and galena in a gangue quartz and adularia, with secondary zeolites.

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[1] R. Beck, Lehre von den Erzlagerstätten, vol. 1, 1909, p. 334.

Genesis.—In the preceding pages attention has been called to the strong evidence connecting the class of deposits here discussed with igneous action and pointing to ascending hot waters as the agents of deposition. The best proof that the ores were not formed by the ordinary circulation of surface waters is the fact that deposition has not proceeded uniformly, but that the vein-forming epochs were of brief duration and followed closely after each considerable eruption. Evidence of this relation is available from many important districts. At Tonopah the principal mineralization followed the eruption of the earlier andesite and the veins are truncated by the flow of the later andesite and the later rhyolite. At Jarbidge, Nevada, the veins are contained in the early rhyolite, while the later rhyolite is barren. At Waihi, New Zealand, the rich veins are sharply truncated by erosion and capped by a rhyolite of later age.

The occurrence of these deposits in lavas really counts for but little; there are vast areas of lava flows absolutely barren of mineral deposits. On the other hand, several of the Hungarian authors have pointed out the fact that the veins are confined mainly to the vicinity of volcanic necks or centers of eruption. Exactly the same conclusions have been reached in the United States. This feature serves to connect the veins formed near the surface with those of greater depths. The deposits in the surface lavas are, then, simply the tops of veins, the roots of which are to be found in the intrusive masses of the depths. No matter whence all the water or part of the water came, the deposition of the substance of the veins—their valuable content—appears to be a phenomenon connected with intrusive activity and not merely dependent upon the heat furnished by the lava flows to circulating surface waters. The metals, as well as the sulphur, carbon dioxide, and fluorine, were in all probability derived from intrusive underlying masses.

Proof of Depth below Surface.—Physiography furnishes the data on the original surface during deposition. We may be able to trace the old surface of the volcanic slope or plateau and ascertain the relation of the outcrops to the uppermost flow, or in dissected volcanic piles it may be possible to reconstruct approximately the old surface of the volcanic cone. Of this latter possibility Cripple Creek is an instance (Fig. 167); the present surface was probably less than 1,500 feet below the original surface of the volcanic cone. Ransome estimates that at Goldfield, Nevada, the surface has been degraded but a few hundred feet below the original contours of the flows. A fine example showing the connection of deposits formed near the surface with those of more deep-seated type is offered by the San Juan region, in Colorado, where erosion has not only intersected the flows but laid bare the intrusive masses forced into them—all within a vertical interval of 6,000 feet.

Proof of Temperature.—The similarity to hot-spring deposits is least marked in deep-seated veins, but becomes striking in the veins here under consideration. The fine-grained chalcedonic and banded quartz of spring deposits (Fig. 5, p. 101) is entirely similar to the often delicate and beautifully banded and crustified portions of these veins. The evidence indicates deposition by waters that held in solution large quantities of substances not easily soluble—that is, by hot waters which at the surface could not have had a temperature of more than 100° C. The minerals present are those which we have reason to believe were developed at a temperature less than 200° C. What the actual temperatures were in each case is of course scarcely possible to ascertain.

The hot springs are volcanic "after effects" and usually ascend through the cooled lavas. In some cases the waters rise through bodies of hot rocks and then the pressure may become so high that the solutions issue at the surface as gases and form "fumaroles" and "soffioni" which sometimes, at their orifices, have a temperature of as much as 200° C. In these rarer instances the high temperature deposits, marked for instance by tourmaline or cassiterite, may develop close to the surface.

Relation to Other Veins.—The question naturally arises as to the character of these veins in depth. Do they actually change to assume the aspect of the veins of the deeper zones? The evidence, scant as it is, indicates that this is probably true. In regions of deeply eroded volcanic flows, like the San Juan country in Colorado, the veins in the lower exposures show an approach to the types of deep-seated origin. During the long ascent there was no doubt a progressive change in the nature of the depositing waters; some of their constituents were deposited and others were acquired from the rocks they traversed.

METASOMATIC PROCESSES

Extent of Alteration.—At considerable depths the ore-forming solutions move in the paths prescribed by fissuring and brecciation; they rarely penetrate great masses of rocks. Near the surface, especially in the great volcanic piles, different conditions prevail. There are here thick beds of tuffs and agglomerates with great porosity, and the stresses may irregularly shatter large volumes of rocks. The solutions and gases—of meteoric or telluric origin—move far more freely and alteration is effected on the largest scale. Here, too, we find most emphasized the peculiar effects of the mingling of ascending and descending solutions.

Any one who has visited an active or recently extinct volcano has undoubtedly observed the areas of discolored reddish, brown, and yellow rocks which indicate alteration. Erosion of older volcanoes discloses similar zones of alteration on a large scale and exposes metalliferous deposits formed in their interior.

Types of Alteration.—Gases given off by the ascending lavas penetrate the volcanic cones. They are admixed with water vapor, which may or may not be of intratelluric origin, and, near the point of issue, oxidation and interaction produce compounds like sulphur dioxide, sulphuric acid, hydrochloric acid, and sulphur. Of these reagents carbon dioxide and sulphuric acid are most effective in rock alteration. The volcanic rocks are converted to kaolin. Alunite, jarosite, and other sulphates are often mixed with these minerals.

These masses of altered rocks, which are formed on the slopes of volcanoes, scarcely ever carry valuable ores, probably because the metallic load of ascending waters is usually deposited before the cool surface zone is reached.

Hot springs begin to issue after the explosive igneous action has declined and the rocks cooled so that the fumarolic action is supplanted by rising aqueous solutions. These waters contain no strong acids, but probably mainly carbon dioxide, silica, and hydrogen sulphide, and are of alkaline rather than acid reaction. Some of these waters move slowly, percolating through great masses of rocks; others move rapidly in prescribed channels and effect extensive changes in the immediately adjoining rock.

One of the most common types of alteration is that resulting in the "propylitic facies;" it affects mainly andesites and basalts, more rarely rhyolites, often spreading over wide areas in mineralized districts. Its mineralogical characteristics consist in the abundant development of chlorite and pyrite, sometimes also epidote; in places it is accompanied by the development of carbonates and a little sericite. The rock assumes a dull green color. The chemical changes consist of a moderate leaching of both potassium and sodium; the silica is usually decreased, as are also calcium and magnesium, except when carbonates of these metals are formed. The composition of the rock changes but little and the additions consist only of sulphur and some water of hydration.

Still another type of alteration, seen mostly in siliceous rocks like rhyolite, but also in other kinds, consist in a general silicification of the groundmass and phenocrysts, with aureoles of quartz developing around quartz phenocrysts. More rarely sericite develops in abundance and the effusive rocks are converted over large areas to a mixture of quartz and sericite, with more or less pyritization.

Near the veins the alteration is usually most intense, although here, too, simply chloritized rock may often adjoin the fissure filling. In sericitization sodium is almost entirely carried away and potassium is accumulated in a marked degree in sericite and adularia; the latter mineral has a wide distribution, both in the altered country rock and in the fissure filling. Unless carbonates are formed, calcium and, to a less degree, magnesium are carried away; much pyrite is introduced which usually contains at least traces of precious metals. The percentage of silica is reduced. Close to the vein, silicification often assumes the ascendancy and a quartzose mass of silica, adularia, and sericite, with more or less sulphides, develops and may form part of the ore. In rare cases hydrargillite and zeolites may appear in the altered rock. Rutile appears to be the only stable titanium mineral. Manganese, titanium, and phosphorus are partly removed from the rock.

Nearer the surface another potassium-aluminum mineral—alunite—appears in considerable quantities. This hydrous sulphate is characteristic of large altered areas in volcanic regions,[1] but, being inconspicuous, is easily overlooked. That it often occurs together with pyrite and sericite is clearly proved, and its development in this phase is probably confined to the zone

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[1] B. S. Butler and H. S. Gale, Alunite, *Bull.* 511, U. S. Geol. Survey, 1912.

where the descending waters carrying free sulphuric acid meet the ascending currents of alkaline waters. It appears to belong to a distinctly higher horizon than the sericite and adularia. In some alunites the potassium is in part replaced by sodium.

In eroded and mineralized volcanic regions there is finally another type of alteration, the effects of which were superimposed upon the earlier changes and tend to confuse the true history. As soon as the mineralized rocks become exposed to the air oxidation begins and sulphuric acid is generated by the action of oxygen on sulphides. This sulphuric acid descends with the surface water and converts the sericitized rocks into kaolin mixed with alunite and other oxidation products. Where waters exceptionally rich in sulphuric acid have acted on the rocks almost everything but quartz is carried away and the final result is a loose quartz aggregate. Descending still farther these sulphuric acid solutions may lose their oxygen, and, under certain circumstances, secondary sulphides, with alunite and sericite, may again develop.

This brief sketch indicates how complicated the series of reactions may be and how the same minerals may form at different steps of the process.

It is assumed in the above discussion that ascending alkaline waters do not form kaolin. This is undoubtedly true in general, but it is possible that kaolin may be formed in places by such waters close to the surface. The processes of alteration by hot ascending waters seem to result in minerals of only moderate hydration; zeolites, kaolin, and other strongly hydrated minerals are conspicuously absent. The zeolites appear to require quiescent, stagnant conditions, such as do not exist close to strong ascending currents.

Metasomatic Processes at Thames and Waihi.—Extensive propylitization has been described by several authors from observations in the Hauraki Peninsula, in the northern island of New Zealand,[1] where rich gold-bearing veins appear in volcanic rocks like pyroxene andesite or dacite. The extreme phase close to the veins is a grayish-white rock, but a more widespread type is a chloritized andesite which corresponds to the propylitic facies as defined on a previous page. In this second type the ferric

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[1] A. M. Finlayson, Problems in the geology of the Hauraki gold fields, *Econ. Geol.*, vol. 4, 1909, pp. 632–645.

minerals are chloritized, the pyroxene often passing first through a uralitic stage, while the plagioclase remains comparatively fresh, but contains some calcite and sericite.

Finlayson has presented two extremely valuable series of analyses, which are given below in full. They represent rocks from Thames and Waihi, the two most important fields in the peninsula. The first column in each table gives the composition of the fresh rock, the second that of the propylitic or chloritic facies, and the rest are analyses of the more altered forms in which sericite and adularia are the predominating metasomatic products. The chemical changes during propylitization are not great.

ANALYSES OF FRESH AND ALTERED ROCKS IN THE THAMES DISTRICT

	1	2	3	4	5
SiO_2	57.42	52.69	57.99	55.38	58.98
TiO_2	0.68	0.53	0.51	0.24	0.11
Al_2O_3	17.61	18.33	17.59	15.63	11.21
Fe_2O_3	2.34	2.32	1.56	1.88	1.45
FeO	3.77	2.98	2.37	2.95	2.42
MnO	0.43	0.25	0.21	0.23	0.11
MgO	2.19	3.09	2.01	1.88	1.43
CaO	5.69	7.87	5.45	6.01	8.11
Na_2O	3.22	2.62	1.98	0.83	0.61
K_2O	1.94	0.98	1.65	3.28	3.93
H_2O-	0.85	0.73	1.56	2.41	2.54
H_2O+	2.62	3.71	1.89	1.92	1.15
CO_2	0.95	3.59	3.89	4.58	4.69
P_2O_5	0.31	0.42	0.35	0.11	0.06
FeS_2	....	....	1.42	2.35	3.13
Total	100.02	100.11	100.43	99.68	99.93

1. Fresh hornblende andesite, Thames.
2. Chloritized hornblende andesite, Halcyon mine.
3. Altered andesite, 14 feet from Ophir vein, Halcyon mine.
4. Altered andesite, 5 feet from Ophir vein, Halcyon mine.
5. Altered andesite, adjoining Ophir vein, Halcyon mine.

(3–5:) Sericite-pyrite-carbonate rock from the 386-foot level.

ANALYSES OF FRESH AND ALTERED ROCKS AT THE WAIHI MINE

	1	2	3	4	5	6
SiO_2	63.45	58.39	61.78	69.35	76.61	85.65
TiO_2	0.75	0.68	0.69	0.43	0.28	tr.
Al_2O_3	15.26	16.51	14.89	11.66	8.31	1.35
Fe_2O_3	2.28	2.46	2.08	1.53	1.08	0.43
FeO	3.01	2.98	2.51	1.66	0.59	0.21
MnO	0.36	0.32	0.28	0.11	0.11	0.12
MgO	1.29	1.66	1.08	0.46	0.51	0.31
CaO	3.44	4.08	3.16	2.09	3.61	2.56
Na_2O	2.21	2.08	2.18	1.06	0.29	0.28
K_2O	1.78	2.89	3.68	3.31	1.98	1.41
H_2O-	1.10	2.41	1.89	1.61	0.43	0.24
H_2O+	2.90	2.87	3.05	2.12	1.08	1.33
CO_2	1.08	1.56	2.01	2.24	1.87	2.04
P_2O_5	0.29	0.31	0.30	0.26	0.11	tr.
FeS_2	....	....	0.65	1.88	3.59	4.69
Total	99.20	99.20	100.23	99.77	100.45	100.62

1. Fresh hornblende dacite, Waihi.
2. Chloritized hornblende dacite, 45 feet from Empire vein.
3. Altered dacite, 30 feet from Empire vein.
4. Altered dacite, 15 feet from Empire vein.
5. Altered dacite, adjoining Empire vein.
6. Replacement ore, Empire vein.

(3–6: 850-foot level.)

Propylitization involves a distinct hydration, caused by the development of chlorite. Where carbonates are formed, magnesia and lime, especially the latter, are somewhat increased. The percentages of alkali metals decreases, but only in moderate degree. If sericite has formed, the potassium may be somewhat higher in the altered rock. Within the influence of the vein-forming solutions the normal alteration to sericite and adularia asserts itself. The two excellent series of analyses quoted above show a slightly differing trend. At Thames the altered rocks contain 10 or 11 per cent. of carbonates, while at Waihi the carbonates form only one-half of that amount. As in the California gold-quartz veins, this development of carbonates results at Thames in the fixing of calcium, while magnesium shows slight changes. At Waihi there is little change in calcium, while the magnesium has been somewhat reduced. In both places there is strong leaching of sodium and progressive accumulation of potassium, except that at Waihi the potassium finally diminishes in the highly quartzose vein material. Iron in ferric and ferrous state is converted to pyrite, but the total iron is not much increased. At Thames, where carbonates are abundant, the silica tends to decrease; at Waihi the opposite is true. In both places there is an apparent decrease in alumina, and also a remarkable and unmistakable leaching of titanium, phosphorus, and manganese, as has also been noted by Spurr at Tonopah.

Mineralogically the alteration near the vein results in sericite, calcite, siderite, pyrite, quartz, and adularia, the last mineral in places forming pseudomorphs after soda-lime feldspars, while it also occurs in small fissures. The adularia (valencianite) from Waihi was analyzed by Finlayson and found to contain 11.25 per cent. K_2O and 4.11 per cent. Na_2O, while the material from Silver City, Idaho, and Tonopah, Nevada, previously examined yielded only a very small quantity of Na_2O.

Stilbite and laumontite have been identified in the altered rocks of Waihi,[1] and analyses 4 (Waihi) and 5 (Thames) suggest the possibility of their presence.

Finlayson does not accept Spurr's view that the vein-forming waters, filtered through rock masses, caused propylitization, but thinks that this alteration is due to solutions or gases rich in CO_2, which permeated the rocks immediately after solidification; the sericite-pyrite carbonate rock along the veins, according to Finlayson, is caused by ascending solutions of a different class.

Metasomatic Processes at Tonopah.—During the alteration of the trachyte near the veins at Tonopah, Nevada,[2] biotite and hornblende, have usually been completely destroyed; their outlines are marked by aggregates of sericite, quartz, pyrite, and siderite, the latter two often crystallizing together. The primary andesine-oligoclase has changed to quartz and sericite or to adularia; the latter two are not often associated in the same specimens. The microlitic groundmass is largely altered to fine-grained quartz with fibers of sericite; pyrite and siderite are disseminated. Apatite and zircon are residual minerals. Kaolin, when present, is believed to result from the alteration of sericite by descending solutions.

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[1] P. G. Morgan, *Trans.*, Aust. Inst. Min. Eng., vol. 8, 1902, p. 186.

[2] J. E. Spurr, Prof. Paper 42, U. S. Geol. Survey, 1905; Econ. Geol., vol. 10, 1915, pp. 713–769.

At a distance from the larger veins a propylitic type of alteration appears, in which calcite and chlorite, together with pyrite and siderite, are the important minerals. The feldspars are altered to calcite with a little quartz; epidote is not abundant. There are transitions between the propylitic and the sericitic alteration, and according to Spurr they were produced by the same waters. Near the veins these waters introduced silica, potassium, and metallic sulphides; as they penetrated farther

ANALYSES OF FRESH AND ALTERED ROCKS, TONOPAH, NEVADA

	1	2	3	4	5
SiO_2	67.69	55.60	72.98	73.50	91.40
TiO_2	....	0.72	0.44	0.47	0.07
Al_2O_3	17.67	16.70	14.66	14.13	4.31
Fe_2O_3	2.43	2.23	1.01	1.51	0.77
FeO	0.80	3.51	0.16	0.26	0.11
MgO	0.88	2.60	0.33	0.21	0.18
CaO	0.45	4.27	0.18	0.12	none
BaO	....	0.12	....	0.19	0.02
Na_2O	2.54	4.08	none	0.24	0.06
K_2O	5.11	3.17	6.03	5.11	1.68
H_2O-	....	0.88	0.97	1.07	0.46
H_2O+	....	3.06	2.95	2.81	0.98
CO_2	....	2.76	none	none	none
P_2O_5	....	0.28	0.16	0.09	0.04
	97.57	99.98	99.87	[1]99.71	[2]100.08

1. Partial analysis of relatively fresh "Mizpah" trachyte. Booth, Garrett and Blair, analyst.

2. Altered andesite, Siebert shaft. Propylitic alteration to quartz, calcite, chlorite, and sericite. George Steiger, analyst.

3. Altered trachyte, Mizpah mine. No original minerals remaining. Sericitic alteration. George Steiger, analyst.

4. Altered trachyte, Mizpah Hill. Typical alteration to adularia with a little sericite. George Steiger, analyst.

5. Ore material of Mizpah vein. Dense quartzose rock mixed with kaolin-like material. Silicified trachyte. George Steiger, analyst.

[1] Also 0.17 SO_3 and 0.03 S.

[2] Also 0.12 ZrO_2 and 0.06 MnO.

from these channels their metal contents were exhausted, while silica and potassium were still introduced; finally only carbon dioxide and hydrogen sulphide were left in the cooling waters, which, therefore, had little to precipitate and small power of abstracting. The wall rock acted as a screen for the traversing solutions.

As noted above, these views are not entirely accepted by Finlayson.

The most prominent features of the alteration, as shown by analyses, are the almost complete removal, adjacent to the veins, of ferrous iron, calcium, magnesium, and sodium and the partial removal of ferric oxide. Even the resistant apatite and rutile seem to have been dissolved to some degree, as shown by the relations of phosphorus and titanium. On the other hand, there is a decided increase of silica, and the potassium has increased. There is a moderate hydration, but no carbonates appear.

A later mineralization, which affected the later andesite at Tonopah, is materially different; the waters by which it was effected appear to have contained practically no gold and silver. The course of the alteration[1] involves no silicification and practically no change in calcium. The sodium is, as before, almost wholly removed, and likewise a large part of the potassium. Carbonates are present in abundance, with pyrite, and some zeolite is probably present, possibly also some talc and hydrargillite.

The Development of Kaolin.—It has been stated above that kaolin in the altered rocks of mineral deposits results mainly from the leaching by surface waters containing free sulphuric acid, and that this mode of alteration is frequently superimposed upon the products of chloritic and sericitic alteration by ascending waters. The sulphuric acid attacks and removes all calcium, magnesium, sodium, and potassium; and the final result is a mixture of kaolin and quartz. Below the influence of free oxygen sulphides may be deposited with the kaolin; pyrite, more frequently marcasite in arborescent forms, chalcocite, covellite and rich silver ores like argentite and stephanite may also develop (see discussion of sulphide enrichment, Chapter 31). In places—for instance, at De Lamar, Idaho—this kaolin may contain much gold in extremely finely divided state, undoubtedly concentrated by secondary reactions.

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[1] *Prof. Paper* 42, U. S. Geol. Survey, 1905, p. 241.

Metasomatic Processes at Silverton, Colorado.—The process of kaolinization is well described in Ransome's report on the Silverton district, Colorado.[1] Crystallized kaolinite—a rare occurrence—was found in the National Belle mine, but is here, too, later than the ore.

The normal alteration at Silverton is of propylitic aspect, changing near the veins to sericitic facies. In a series of rocks occurring in the Silver Lake basin the andesite breccia 150 feet from the vein is only slightly altered by the destruction of the dark silicates and by the beginning of replacement of feldspars by sericite (perhaps with some kaolin) and calcite. At 100 feet from the vein the quantity of chlorite and calcite increases. Fifty feet away from the vein the breccia structure is still visible, but the rock is wholly re-crystallized to quartz, chlorite, sericite, calcite, and rutile, with residual apatite. Two feet from the vein there is less chlorite, and the rock consists mainly of sericite and quartz, with some grains of galena. Close to the wall there is but little chlorite, and considerable pyrite has been introduced. This general process corresponds fairly closely to that at Tonopah.

In the same region, at Red Mountain, the alteration of the rocks is carried to its ultimate conclusion. The mine waters show the presence of free sulphuric acid and alumina and the white kaolinized rock at the surface shows the following composition calculated from the analysis.

COMPOSITION OF ALTERED ROCK AT RED MOUNTAIN, NEAR NATIONAL BELLE MINE

Quartz	60.9
Kaolinite	26.3
Pyrite	5.6
Diaspore	3.8
Sericite	0.6
Apatite	0.6
Rutile	0.6
	98.4

A still more advanced silicification is shown by the following analysis of an altered rock at the White Cloud Mine in the same region, from which the mineral composition of the rock may be calculated as follows:

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[1] F. L. Ransome, Economic geology of the Silverton quadrangle, *Bull.* 182, U. S. Geol. Survey, 1901.

Quartz	78.5
Kaolin minerals	16.0
Pyrite	3.4
Rutile	0.6
Sulphates (Fe, Ca, Ba)	1.3
	99 8

This extreme mode of alteration by sulphuric acid solutions results in the almost total elimination of calcium, magnesium, alkali metals, and carbon dioxide; its operation is similar to that of weathering by oxygenated waters, although in that process, of course, pyrite cannot form. It differs radically from the sericitization and carbonatization described above.

Summary.—Summing up we may say that sericite and carbonates mark the alteration by ascending metalliferous solutions at intermediate and moderate depths; that nearer the surface adularia appears as an important metasomatic mineral in addition to sericite; and that still nearer the surface or under the influence of descending sulphuric acid solutions we find kaolin, or, where sulphuric acid is present in abundance, alunite.

QUICKSILVER DEPOSITS

The Ores and Their General Occurrence.—The principal quicksilver ore is cinnabar (HgS), which contains 86.2 per cent. mercury. A black modification of this mineral, called metacinnabar, rarely occurs in large amounts and is probably a secondary sulphide deposited by descending waters.[1] Native quicksilver, silver and gold amalgam, calomel (HgCl), quicksilver oxide (montroydite), and several oxychlorides are evidently secondary minerals, developing from the sulphide (p. 892). Primary but rare minerals are the black telluride of quicksilver, coloradoite; the selenide, tiemannite; the sulphoselenide, onofrite; and other still rarer combinations of the selenides of copper, lead, and quicksilver. Mercurial tetrahedrite is not uncommon, and some varieties contain as much as 17 per cent. quicksilver, although the percentage is usually much smaller. In smaller quantities this metal is also sometimes present in other minerals, for instance, in the native silver of Kongsberg, Norway, and in the dyscrasite of the silver-bearing veins of Cobalt, Ontario (p. 626).

The occurrence of quicksilver minerals is by no means confined

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[1] E. T. Allen and J. L. Crenshaw, The sulphides of zinc, cadmium and mercury, *Am. Jour. Sci.*, 4th ser., vol. 34, 1912, pp. 367–383.

to any certain kind of deposits or to any given age or epoch of metallization. However, such minerals are not known to occur in deposits of distinctly igneous origin, nor in pegmatite dikes, nor in veins of the deepest zone. High temperature is evidently unfavorable for their development. The most noteworthy occurrence is that of coloradoite in the gold telluride veins of western Australia, which contain, among other minerals, magnetite and tourmaline, indicating deposition at fairly high temperature. In gold-bearing quartz veins of the ordinary type, believed to have been formed at a considerable depth, but at considerably lower temperature and pressure than pegmatite dikes, cinnabar is not an uncommon mineral. It occurs in several of these veins in California,[1] as well as in the similar veins of central Idaho, and is frequently found in the placers derived from the erosion of these veins, as at Stanley Basin and Warren, Idaho. In northeastern Oregon the gold-quartz veins contain mercurial tetrahedrite, as well as secondary cinnabar formed from that mineral.[2] In the placers below the veins of Susanville, in the same region, pebbles showing masses of cinnabar inclosed in massive white vein quartz have been found. One often finds apparently reliable statements that during the process of amalgamation and refining of the gold from such deposits more quicksilver was recovered than was added for metallurgical purposes. In small quantities cinnabar occurs in the lead and zinc deposits of Monteponi, in Italy, and at Santander, in Spain. Many occurrences of mercurial tetrahedrite in Europe and South America have been described.

In few of these deposits are the mercurial minerals abundant enough to constitute an ore, and in the majority of the deposits formed at a considerable depth the metal is apparently entirely absent. The commercial quicksilver ores are practically confined to a small and well-defined group of deposits, which will be described in the following pages and which are of particular interest because their genesis can be fairly accurately ascertained.

A scant association of ore minerals characterizes these deposits. Besides cinnabar and metacinnabar, as well as a few minerals derived from the decomposition of the sulphide, they contain

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[1] H. W. Turner, *Am. Jour. Sci.*, 3d ser., vol. 47, 1894, p. 467.
H. D. McCaskey, *Min. Res.*, U. S. Geol. Survey, pt. 1, 1910, p. 905.

[2] W. Lindgren, The gold belt of the Blue Mountains of Oregon, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, pp. 604, 708.

almost invariably pyrite or marcasite and frequently stibnite, but rarely any of the sulphides of the base metals so common in ore deposits. Among gangue minerals we have predominatingly opal, chalcedony, and quartz, also calcite and dolomite, more rarely barite, and very seldom fluorite; zeolites are of exceptional occurrence.

As to form and structure the ores occur in irregular and "chambered" veins and brecciated zones, often also as "stockworks" of minute seams, or as disseminations in more or less porous rocks. The irregularity and brecciated character of the deposits suggest their development near the surface, a conclusion that is often justified by other geological evidence.

As to association the deposits occur in rocks of any kind and any age, but almost always in close connection with effusive rocks or in regions of volcanic activity. Hot springs are frequent in many quicksilver districts, and this conspicuous association has led to the view that the metal is a last product of differentiation in many magmas nearing the surface and that the hot springs take it into solution to deposit it near their points of issue. This theory is strongly fortified by the discovery that many springs in volcanic regions are depositing cinnabar at the present time.

Although most of the quicksilver deposits have been formed at a relatively late time and in connection with the eruption of Tertiary and recent lavas, it does not necessarily follow that their development has been confined to late geological time. Older surface eruptions were undoubtedly accompanied in places by the formation of quicksilver deposits, but as these were near the surface they were easily eroded. Attention has already been called to this elsewhere,[1] and it has been suggested that the deposits at Almaden, in Spain, and those in the German Palatinate may belong to such more remote volcanic epochs.

Distribution, Production and Use.—Quicksilver deposits are widely distributed, although the main production of the metal comes from a few occurrences. G. F. Becker, who first studied the occurrence of these ores concluded that their distribution follows the main structural lines of the continents, especially in the Pacific region and in the Alpine-Himalayan chain. While this is apparently true, it is more correct to say that the ores follow the belts of Tertiary and Quaternary eruptions, especially along the important fracture lines of the globe.

The Coast Range of California, in which the orogenic move-

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[1] Beyschlag, Krusch, and Vogt, Die Lagerstätten, etc., vol. 1, 1910, p. 454.

ments are largely post-Miocene, contain a belt of quicksilver deposits several hundred miles in length, from which at one time a large production was obtained.[1] A second belt, perhaps less well defined and certainly less productive, extends from north to south over a similar length in western Nevada.

The Mexican area, which in spite of comprising many deposits, yields only a slight production, begins in western Texas in the Terlingua district[2] and may be considered to end in the State of San Luis Potosi, Mexico.[3]

Farther south, in Peru,[4] quicksilver deposits appear again. The Yauli and Huancavelica districts are best known; the latter at one time was an important producer.

In Europe an extensive region in Italy, Austria, and adjacent countries contains quicksilver deposits; this area includes the deposits of Tuscany,[5] Vallalta-Sagron,[6] Idria and vicinity,[7]

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[1] G. F. Becker, The quicksilver deposits of the Pacific slope, U. S. Geol. Survey, *Mon.* 13, 1888.

W. Forstner, The quicksilver resources of California, *Bull.* 27, California State Min. Bur., 1903.

H. D. McCaskey, *Mineral Resources*, U. S. Geol. Survey, pt. 1, 1910–1916, particularly in issue of 1911, with literature.

[2] W. B. Phillips, The quicksilver deposits of Brewster County, Texas, *Econ. Geol.*, vol. 1, 1906, pp. 155–162.

H. W. Turner, The Terlingua quicksilver deposits, *Econ. Geol.*, vol. 1, 1906, pp. 265–281.

[3] J. D. Villarello, Génesis de los yacimientos mercuriales de Palomas y Huitzuco, *Mem.* Soc. Ant. Alzate, vol. 20, 1903, pp. 95–136.

J. D. Villarello, Descripción de los criaderos de mercurio de Chiquilistan, Jalisco, *Idem*, vol. 20, 1904, pp. 389–397.

P. A. Babb, Dulces Nombres quicksilver deposit, Mexico, *Eng. and Min. Jour.*, Oct. 2, 1909.

F. J. H. Merrill, The mercury deposits of Mexico, *Mining World*, vol. 24, 1906, p. 244.

[4] A. F. Umlauff, El cinabrio de Huancavelica, *Bol.* 17, Cuerpo de Ingenieros de Minas, Lima, 1904, p. 61.

[5] V. Spirek, Die Zinnobererzvorkommen am Monte Amiata, *Zeitschr. prakt. Geol.*, 1897, pp. 369–374; *idem*, 1902, pp. 297–299.

B. Lotti, Il campo cinabrifero dell' Abbadia, etc., *Rass. Min.*, vol. 7, 1898, No. 11; *Zeitschr. prakt. Geol.*, 1898, p. 258. See also *Rass. Min.*, vol. 17, 1902, No. 10.

[6] A. Rzehak, Die Zinnoberlagerstätte von Vallalta-Sagron, *Zeitschr. prakt. Geol.*, 1905, pp. 325–330.

[7] F. Kossmat, Ueber die geologischen Verhältnisse des Bergbaugebietes von Idria, *Jahrb.* K. k. geol. Reichsanstalt, vol. 49, 1899, pp. 259–286.

Geologisch-bergmännische Karten, etc., von Idria. Text by Plaminek. Published by the Agricultural Department, Vienna, 1893. (Literature.)

Avala,[1] in Servia, and less important occurrences in Bosnia.

Isolated yet highly productive deposits occur in Almaden, in Spain.[2]

Some deposits have been found on the western side of the Pacific, mainly in Japan, China, Borneo, Australia, and New Zealand.

Before the war, in 1913, the world's production of quicksilver was 124,654 flasks at 75 pounds. Of this Spain produced 43,799, Italy 29,513, Austria 26,720 and United States 20,213. In 1917, the domestic production was 36,315 flasks. California yielded 24,251 flasks. The average price in 1913 was $40 per flask. In 1917 the average for the year was $106. The principal use of quicksilver is for gold amalgamation, drugs, paints and mercurial fulminate ($Hg(ONC)_2$) an explosive used for priming shells.

Geological Features.—The comparative youth of the deposits is attested by the fact that many of them are found in sedimentary or volcanic rocks of Tertiary or Quaternary age. They are not confined to these rocks, however, and may, in fact, occur in rocks of any composition or age. Sandstones, shales, limestone, serpentine, granite, andesite, rhyolite, or basalt may harbor the ores, and the character of the surrounding rocks seems to have little influence on the value of the deposits.

The California belt contains ores in Jurassic, Cretaceous, and Tertiary sandstones and shale, in serpentine, and in late Tertiary or Quaternary basalt and andesite. In the Nevada belt the ores occur in Triassic strata in Paleozoic limestone and dolomite or more commonly in rhyolite, probably of middle Tertiary age.

In Texas and Mexico the ore-bodies are in Cretaceous strata or in the Tertiary andesite, basalt, and rhyolite which break through them. The Peruvian deposits are in Jurassic beds or in Tertiary volcanic rocks.

In the Adriatic region of Europe the ores occur in rocks of many kinds: In Tuscany, Mesozoic and Tertiary limestones and sandstones with trachyte are the enclosing rocks; at Idria, the

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[1] H. Fischer, Die Quecksilberlagerstätten am Avala-Berge in Serbien, *Zeitschr. prakt. Geol.*, 1906, p. 245.

[2] There is no modern and detailed description of this important deposit. The best account is found in Beck's "Lehre von den Erzlagerstätten," vol. 1, 1909, pp. 519–522.

disturbed beds of the Alpine Triassic; at Avala, the serpentine and probably Cretaceous limestone cut by trachytic dikes. De Launay[1] has shown that these Adriatic ores coincide in their extension with Tertiary eruptives, and that in all probability, even where these eruptives are locally absent, as at New Idria, the deposits owe their origin to the after-effects of this igneous activity in the form of ascending springs.

In the Donetz basin in southern Russia[2] the cinnabar ores lie in Carboniferous strata and have no apparent connection with igneous rocks.

Mineralogy of Quicksilver Ores.—Cinnabar while usually massive sometimes forms well-defined but small crystals; these are especially common in porous rocks like sandstone and tuff.

Aside from pyrite and the more common marcasite the ore mineral most generally accompanying cinnabar is stibnite; many stibnite veins, it should be added, also contain some cinnabar. Quartz is usually present, but far more commonly the silica appears as chalcedony or opal. In the California occurrences opal is particularly abundant and here, as at Avala, much of it is a product of the replacement of serpentine. In California the cinnabar is not often found in the opal itself, but rather in the veinlets of quartz or chalcedony traversing it. Calcite is not an uncommon gangue material, and in the Coast Ranges of California many of the deposits are accompanied by calcium-magnesium carbonates derived by replacement from serpentine or allied rocks.

Among the sulphates barite is fairly abundant, and at most places there is also more or less gypsum, which may often be a product of primary deposition, although it would naturally also be generated by the effect of decomposition of pyrite in a calcareous gangue. Fluorite is rare, but is recorded from Guadalcazar and Idria. In many deposits, particularly those of Idria and the California belt, hydrocarbons are characteristic; they are probably derived from the adjacent sedimentary beds, but are believed to have exerted some influence in the precipitation of cinnabar from its solutions. Inflammable gases, mainly hydro-

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[1] L. De Launay, La métallogénie de l'Italie, International Geological Congress, Mexico, 1906.

[2] *Zeitschr. prakt. Geol.*, 1894, p. 427, after Kulibin. Tschernyschew and Loutouguin, Guide des excursions du VII Congrès géol. internat., No. 16. 1897, pp. 36–45.

carbons, are reported from several localities in California, notably New Idria.

Zeolites are not unknown in quicksilver deposits; chabazite colored by cinnabar is mentioned from Almaden, in Spain, and apophyllite from New Almaden, in California.

Alunite has been found associated with opal and cinnabar in rhyolites of Nye County, Nevada, near Bullfrog.[1]

Millerite, or sulphide of nickel, is not uncommon in the California occurrences; this metal is probably derived from adjacent bodies of serpentine or peridotite.

The rare tiemannite and onofrite were at one time obtained near Marysvale, in southern Utah, and some quicksilver was recovered from such ore. The minerals occurred as fissure filling in limestone, but in a district of volcanism with rhyolitic and andesitic flows. Selenides are also reported from Guadalcazar, in San Luis Potosi, Mexico. It is probable that careful examination would disclose the presence of selenides at many other places. The mineral livingstonite ($HgSb_4S_7$) occurred in considerable quantities at Huitzuco, Guerrero, Mexico. Regarding oxidation of quicksilver ores see p. 892.

Quicksilver ores should contain at least 0.5 per cent. of the metal. The richest ores are those of Almaden, Spain, which are said to average 8 per cent., while at New Idria ores are mined which contain less than 0.5 per cent.

Structure.—In their structural relations the majority of quicksilver deposits clearly indicate their origin near the surface. Sharply defined continuous fissure veins occur only exceptionally; far more common are irregular and "chambered" veins—that is, fissures accompanied by brecciated masses in which the ore minerals have lodged (Fig. 150). Another common mode of occurrence is as disseminations in porous rocks, like sandstone and tuffs, or again as "stockworks," the ore minerals filling little crevices and fissures in limestone, serpentine, or other rocks.

Cinnabar is often deposited in open cavities or in pores or in the soft mud of altered rocks. It does not replace limestone on the scale of the lead deposits, but that replacement has occasionally occurred seems to be beyond doubt.

A large number of deposits have been found to cease or become impoverished at a depth of a few hundred feet. In contrast to

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[1] Adolph Knopf, Some cinnabar deposits of western Nevada, *Bull.* 620, U. S. Geol. Survey, 1915, pp. 59–68.

this, the celebrated mines of Almaden, Spain, are said to have found richer ore in depth, and the workings have now attained a depth of 1,300 feet.

The deposits at Almaden occur in three beds of steeply dipping Silurian quartzite separated by bituminous slates. In part the cinnabar may occur as filling of the pores of the rock, as G. F. Becker[1] suggests, but Beck[2] has shown convincingly that there has also occurred an actual replacement of the sandstone grains by the ore mineral (Fig. 151). The ore-bodies are as much as 45 feet in thickness, and the average tenor of the ores is unusually high; they are said to contain 8 per cent. quicksilver. Granite and diabase break through the sedimentary series, but the geological history of the deposit is too imperfectly known to draw safe conclusions as to its age or mode of origin. Almaden is the richest and most productive quicksilver region in the world. The value of its metallic product, according to the handbook of Beyschlag, Krusch, and Vogt, during the period 1564 to 1907, is estimated at 212 million dollars.

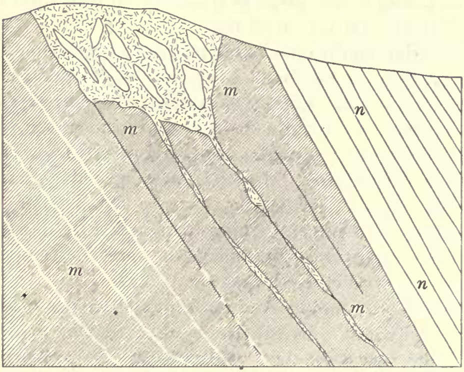

Fig. 150.—Diagrammatic vertical cross section of the Redington cinnabar mine, California, showing brecciated ore chambers near surface, changing in depth to more regular fissures filled with cinnabar. Total depth about 600 feet. *m*, Metamorphic rock; *n*, Cretaceous sandstone. *After G. F. Becker, U. S. Geol. Survey.*

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[1] G. F. Becker, *Mon.* 13, U. S. Geol. Survey, 1888, p. 399.

[2] R. Beck, Lehre von den Erzlagerstätten, vol. 1, 1909, p. 521.

At Idria, in southern Austria, is located another of the great quicksilver mines of the world. The ores are contained in Triassic beds of shale, marl, and dolomite; they are apparently connected with and in part occur in great overthrusts and faults. The ore-bodies, which apparently do not extend below a depth of 1,000 feet, in places follow the stratification and were formerly believed to be of syngenetic origin. The ores are usually designated as "impregnations" in shale or marls, but small veins and stockworks are also found, especially in the dolomite. The age of deposition is certainly post-Cretaceous, probably Tertiary.

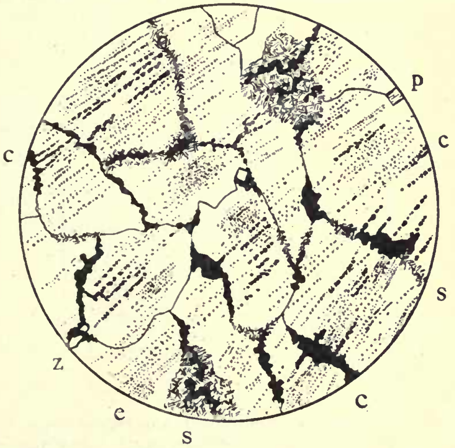

FIG. 151.—Rich ore, Almaden, Spain, showing cinnabar between grains of quartzite and formed by replacement in quartz. c, Cinnabar; s, sericite; p, pyrite; z, zircon. Magnified 70 diameters. *After R. Beck.*

Schrauf, who has given long study to Idria, believes that the ore occurring in the dolomite is a later migration from the somewhat older deposit in the shales.

The California region offers types of almost all the various structural developments. In the region north of San Francisco, near Clear Lake, serpentine, radiolarian cherts, altered Franciscan sandstone (Jurassic?), and Cretaceous sandstones prevail; the rocks are greatly shattered and late Tertiary to Quaternary andesites and basalts break through them. The occurrences of cinnabar are numerous, and some of them have yielded a large production, but the ore-bodies generally become impoverished at a depth of a few hundred feet. The deposits form fissure veins largely filled with attrition material, and this is impregnated with cinnabar, pyrite, opaline silica, and calcite. Masses of ore often extend into the country rock from these fissures (Fig. 152). Or again, as in the Great Eastern mine, the ore forms tabular masses between serpentine and sandstone, or pipes in opaline or chalcedonic rocks between the same formations (Fig. 153), or finally it may be developed on the contact of basalt and sandstone. The Redington mine (Fig. 150) was operated on a large chambered deposit at the surface which was found to be continued below by more regular and narrow veins. Throughout this region hot springs are found in and around the ore deposits.

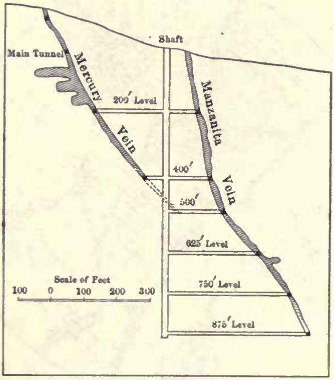

Fig. 152.—Vertical cross section through workings of Napa Consolidated mine, California. Irregular veins in horizontal Cretaceous sandstone, widening to chambers along bedding planes. *After G. F. Becker, U. S. Geol. Survey.*

The great mine of New Almaden, in Santa Clara County, south of San Francisco, is opened in shattered greenstone, serpentine, radiolarian chert, and sandstone of the Franciscan series. Considered in detail the ore-bodies are stockworks, but they are arranged along definite fissures and have on the whole a vein-like character. There are two main fissures of varying dip along and from which the ore-bodies extend. The hanging wall is usually an impermeable, slickensided clay. There are no hot springs and no eruptives in the vicinity. The mine has been opened to a depth of 2,100 feet and a continuous ore-body extended down to the 1,600-foot level. During the last few years little work has been done in the lower levels.

At New Idria, at the south end of the Mt. Diablo Range, important deposits have been worked and the larger part of the production of California is now derived from this mine. The rocks are disturbed greenstones and sandstones of the Franciscan series, unconformably covered by tilted Chico (Cretaceous) and Tejon (Eocene) sandstone. The ores appear in three forms—as normal veins, as irregular stockworks, and as impregnations in sandstone. The mine is opened by tunnels, the lowest level being at a vertical depth of 1,060 feet. There are no volcanic rocks in the immediate vicinity.

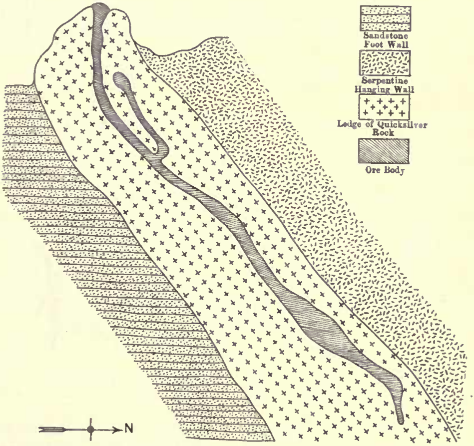

FIG. 153.—Vertical cross section of the Great Eastern mine, California, showing pipe of cinnabar contained in opaline replacement gangue. *After G. F. Becker, U. S. Geol. Survey.*

In the Terlingua district, Texas, near the Mexican boundary, the ores are found in the Upper Cretaceous shales and the Lower Cretaceous limestone. Volcanic rocks are represented by sheets, dikes, and flows of andesite, rhyolite, and basalt. In the lower limestones the ores are mainly in nearly vertical calcite veins, or in lodes of friction breccia (Fig. 154). The other associated minerals are chalcedony, gypsum, aragonite and pyrite.

Genesis.—The uniform character of the quicksilver deposits points to a common genesis for all of them. The earlier belief that the ores were products of sublimation is generally abandoned, for the usual mode of occurrence, with minerals of aqueous origin, such as calcite, opal, chalcedony, and often barite, is decidedly opposed to such a view. Becker has pointed out that, as the character of the enclosing rocks has little influence on the deposits they are most probably derived from a common, deep-seated source. Their structure indicates deposition near the surface, as does also the physiographic evidence at many places—for instance, where the ore appears in the crevices of Quaternary and little-eroded lava flows.

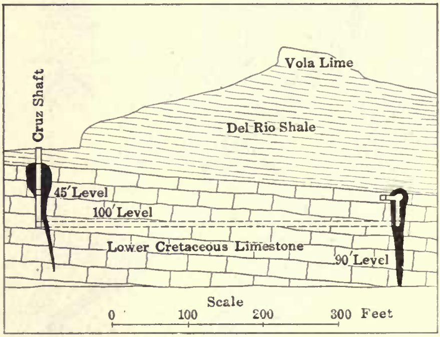

FIG. 154.—Vertical cross section of California Hill, Terlingua, Texas, showing cinnabar veins with large ore-bodies below impervious shale. *After H. W. Turner.*

When it is noted that hot springs and volcanic surface flows are present in almost all regions of importance (except Almaden in Spain, Idria in Austria, and Nikitowka in Russia), and that cinnabar in considerable quantities is associated with hot spring deposits, or is actually found deposited by hot springs, the argument becomes very strong indeed that such solutions have formed the majority of the deposits. For the few deposits that have no such clear connection with volcanic rocks the characteristic mineral association still holds good, and we are forced to the hypothesis that volcanism and hot-spring action are the causes of these also, thoughthe products of the igneous activity may have failed to reach the surface and the hot springs may have subsided.

The evidence relating to cinnabar deposited by hot springs is summarized in the following paragraphs.

At Steamboat Springs, in Nevada, near the California boundary, cinnabar is contained in the hot ascending sodium chloride waters, together with antimony, arsenic, and sulphur, and is actually being deposited in the sinter.[1] Close by, but at a higher level, is a low-grade quicksilver deposit in decomposed granite, and this in all probability was also formed by the same springs when issuing at a higher level. Underneath the sinters of the present springs the gravels contain crystallized stibnite and pyrite.

At Sulphur Bank,[2] in the California quicksilver belt, Le Conte, Christy, Rising, Becker, and Posepny have studied the deposition of cinnabar and sulphur by ascending hot sodium carbonate and borate waters and have all arrived at the conclusion that such deposition, together with that of pyrite and opal, is actually taking place. The Cretaceous sandstones and associated Franciscan metamorphic rocks are here overlain by flows of both normal and glassy basalt and by cinder cones, pointing to very recent eruption. The hot springs have altered and bleached the basalt. Sulphur is deposited at the surface by the oxidation of H_2S, or by reaction between SO_2 and H_2S. Below the superficial deposit of sulphur cinnabar is found in the basalt, as well as in the underlying shales and sandstones; it occurs mostly in veinlets and joints together with the pyrite and opal above mentioned. (Cfr. p. 113.)

The Rabbit Hole sulphur deposit, in Humboldt County, Nevada, described by G. I. Adams,[3] is evidently a product of hot springs, and near it are considerable areas of rhyolite. The rocks are silicified, and opal, alunite, gypsum, and some cinnabar are present as associated minerals.

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[1] G. F. Becker, *Mon.* 13, U. S. Geol. Survey, 1888, Chapter XI.

[2] J. LeConte and W. B. Rising, The phenomena of metalliferous vein formation now in progress at Sulphur Bank, Cal., *Am. Jour. Sci.*, 3d ser., vol. 24, 1882, pp. 23–33.

G. F. Becker, *op. cit.*, Chapter VII.

F. Posepny, The genesis of ore deposits, 2d ed., 1902, pp. 32–36.

[3] *Bull.* 225, U. S. Geol. Survey, 1904, pp. 497–502.

In the Hauraki peninsula of New Zealand, near Omapere Lake,[1] where basalts overlie Mesozoic shales and sandstones, mercury and cinnabar have been found in the deposits of the hot springs at several places.

E. Cortese[2] reports the occurrence of cinnabar in connection with sulphur deposits which result from still active hot springs in Chaguarama Valley, Venezuela. The cinnabar occurs in Tertiary bleached sandstone, together with pyrite. Borax deposits are also said to occur in the same locality.

A careful investigation would doubtless disclose the presence of cinnabar in many other spring deposits. If it is found in more than traces the best way to test such material, as well as ores, is by the miner's pan, in which the bright-red grains of cinnabar show conspicuously.

Quicksilver is apparently contained in hot-spring waters carrying sodium carbonate, sodium chloride, or sodium borate; sometimes all three salts as well as carbon dioxide and some hydrogen sulphide are present. Near the surface these springs may become acid owing to the oxidation of hydrogen sulphide.

Regarding the mode in which mercury is carried in solution, Becker's views,[3] based on the laboratory experiments of W. H. Melville, still appear to furnish the best explanation. While the solubility of mercuric sulphide in alkaline compounds containing sulphur had long been recognized, the evidence was to some degree conflicting.

Becker showed that mercuric sulphide is freely soluble in solutions of sodium sulphide, as well as in a mixture of Na_2S and NaOH, and also in warm sodium sulphydrate (NaHS). When neutral sodium carbonate is treated with hydrogen sulphide, sodium sulphydrate and probably also sodium sulphide will form; these dissolve mercurial sulphide, and double salts of the general formula $HgS.nNa_2S$ doubtless form. Incidentally it was found that the same reagents would dissolve metallic gold, pyrite, sphalerite, and cupric sulphide. The solubility of the

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[1] I. M. Bell and E. de C. Clarke, *Bull.* 8, New Zealand Geol. Survey, 1909, p. 87.

A. Liversidge, *Jour.* Roy. Soc. N. S. W., vol. 2, p. 262.

J. Park, *Trans.*, N. Z. Inst. Min. Eng., vol. 38, 1904, p. 27.

Andre P. Griffiths, The Ohaeawai quicksilver deposits, *Trans.*, N. Z Inst. Min. Eng., vol. 2, 1898, p. 48.

[2] *Eng. and Min. Jour.*, Nov. 10, 1904.

[3] G. F. Becker, *Mon.* 13, U. S. Geol. Survey, 1888, Chapter XV.

sulphides of arsenic and antimony under similar conditions is well known.

It is therefore easy to perceive that a spring water containing sodium carbonate and hydrogen sulphide would form a suitable solvent for the compounds mentioned. The precipitation would be easily effected by oxidation of the water and the development of free acids, by dilution, by cooling, or by the presence of organic or ammoniacal compounds. That the latter two agents are active in many cases there is little doubt.

Relation to Other Ore Deposits.—Although the cinnabar deposits form a well-defined group, they are not to be separated entirely from other classes of ore deposits. Some of them contain other metallic minerals, and there are many that show a transition to the stibnite and arsenical veins. The Nevada belt especially furnishes many instances of a close relationship to gold and silver bearing veins on one hand and to stibnite veins on the other hand. It is true, however, that no cinnabar deposit has yet been found to change gradually into ores of different character as depth is attained. No deposits have been worked below a depth of 2,000 feet vertically beneath the croppings. There is little doubt that these ores were formed from the most volatile parts of the magmas, carried in solution by ascending waters until they came close to the surface. But the irregular distribution of the deposits and their entire absence from large areas of volcanism form problems that are yet unsolved.

STIBNITE DEPOSITS

Mineralogy, Production and Uses.—Stibnite (Sb_2S_3 with 71.4 per cent. Sb) is the principal ore mineral of antimony. Its oxidation near the surface results in various oxides (p. 899) of yellowish or white color such as senarmontite (Sb_2O_3), cervantite (SbO_2), and stibiconite ($H_2Sb_2O_5$). While stibnite occurs in many deposits in small quantities, especially in quicksilver ores, it is the characteristic and dominant mineral in the stibnite veins where it is accompanied by quartz gangue and a scant amount of other sulphides, such as arsenopyrite, realgar, pyrite and zincblende, more rarely jamesonite and similar sulphantimonides. Such ores often also carry gold and the association of the stibnite with some gold quartz veins has often been noted. For the purpose of making pure antimony the presence of arsenic and copper is undesirable.

In past years the production of antimony has not been great owing to its low price and limited usefulness. Since the great war began the price has increased to 17 cents per pound. The world's production of antimony metal may be estimated to 50,000 metric tons. The supply is mainly obtained from China, from Central France, and from the State of Queretaro, Mexico. In the United States but little pure antimony is produced though under stress of war the production of such ores has risen to several thousand tons. Antimony is used for bearing metals, type alloys, shrapnel bullets and its salts find a varied use in the industries. The sulphide is used in pyrotechnics.

Another source of antimony is in the replacement deposits and veins containing mainly galena but associated with tetrahedrite and more rarely with jamesonite, bournonite, boulangerite and other lead sulphantimonides. As a rule these are related to intrusive action and the small amount of antimony contained is recovered as "hard lead" or antimonial lead in the smelting operations. From 2,000 to 3,000 tons of this alloy is produced annually in the United States.

No antimonial mineral is known to occur in magmatic deposits; they are certainly rare in the contact-metamorphic and other high temperature deposits though in these jamesonite, tetrahedrite and stibnite have been occasionally reported (p. 738).

Occurrence.—The stibnite veins have wide distribution but are rarely rich. They are in part formed near the surface, but many deposits are of more deepseated origin and occur in or near intrusive rocks. To the former type belong the stibnite veins with a gangue of fine-grained and drusy quartz which intersect flows of rhyolite and basalt in western Nevada. The antimony sulphide is as a rule beautifully crystallized in acicular and prismatic forms; it is often accompanied by a little pyrite, zinc blende, and arsenopyrite, sometimes also by tetrahedrite and cinnabar. Such veins carry a little silver and less gold. The intimate relationship of these veins with the gold and silver veins proper is, however, shown by the occurrence in one of them, at National, Nevada,[1] of a shoot of remarkably coarse gold of the electrum variety.

Stibnite veins of uncertain affiliations are found in central Arkansas but are of no great importance. The veins follow the steep stratification of Carboniferous shale and sandstone, and

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[1] W. Lindgren, *Bull.* 601, U. S. Geol. Survey, 1915.

the stibnite fills the spaces between the quartz combs.[1] Stibnite shows a marked tendency to form replacements in limestone and shale. Such deposits in Eocene shale below a thick series of andesites have been described from southern Utah.[2] They are undoubtedly hot spring deposits. Of such nature are also the deposits at Pereta, in Tuscany, where the mineral is associated with realgar and cinnabar and occupies veins, seams, and irregular pockets in Tertiary calcareous and detrital rocks. The country rock is in part silicified, in part altered to gypsum or alunite, and exhalations of hydrogen sulphide testify to the recent age of the deposit.

Beck[3] describes important deposits of stibnite at Kostainik, in Serbia, where the mineral occurs in nests and veins in trachyte or in Triassic clay shales, but also as metasomatic bodies replacing the beds along the contact of limestone and shale. The gangue is a drusy fine-grained quartz.

The stibnite veins of Japan, renowned for their beautiful crystals, are found in Mesozoic and Paleozoic rocks but little is known about their affiliations.

The deposit of Djebel Hamimat,[4] in Algeria, containing senarmontite, and that of Altar,[5] Sonora, from which stibnite is reported as the principal ore mineral, appear both to be replacement deposits in limestone. At the Algerian locality the replacement veins spread out in Cretaceous sediments, while at Altar the ore is said to occur in Carboniferous limestone. Both deposits are probably oxidized replacements of stibnite.

Stibnite veins affiliated with intrusive rocks differ but little from the deposits described above. They are known from central France, where narrow veins intersect granite and surrounding schist. Similar deposits are not uncommon elsewhere for instance, in Kern County, California, where the quartz veins also contain gold, in the Coeur d' Alene district and elsewhere.

Stibnite is very common in Alaska and generally is found in gold-quartz veins. A. H. Brooks[6] who enumerates 67 occurrences suggests that the stibnite may have been introduced in older gold-quartz veins during a later and Tertiary mineralization.

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[1] F. L. Hess, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 241–256.

[2] G. B. Richardson, *idem*, pp. 253–256.

[3] R. Beck (after W. von Fircks), *Zeitschr. prakt. Geol.*, 1900, pp. 33–36.

[4] L. De Launay, Gîtes minéraux, 1, 1913, p. 772.

[5] E. T. Cox, *Am. Jour. Sci.*, 3d ser., vol. 20, 1880, pp. 421–423.

[6] Antimony deposits of Alaska, *Bull.* 649, U. S. Geol. Survey, 1916.

GOLD-QUARTZ VEINS IN ANDESITE

Transylvania.[1]—In northwestern Hungary and in adjoining parts of Transylvania gold-bearing veins of Tertiary age have been developed after eruptions of andesites and dacites. A mining industry, begun centuries ago, still flourishes in this region. The literature is very extensive, and only a few deposits can be mentioned here as examples.

The geological formations in the western part of the gold-mining region of Transylvania consist of Mesozoic melaphyres, Cretaceous shales and sandstones, and Miocene sediments, all penetrated by late Tertiary eruptions of andesites and dacites. The igneous rocks appear as lava flows, tuffs, and volcanic necks. The veins near Brád, at present the most productive district, fill well-defined steeply dipping, in places branching fissures which generally intersect volcanic rocks or Cretaceous sediments. The simple veins are as much as 1 meter in thickness; the lodes attain a thickness of 10 to 20 meters. The deposits have been worked to a depth of about 270 meters. The fissures are tectonic, not contraction joints. They are of Miocene age.

The surrounding rocks, particularly the andesite, have suffered extensive propylitization, the femic minerals being extensively decomposed, while feldspars remain fresh. Pyrite is not common except near the veins. Calcite is abundant. Schumacher does not believe that propylitization is caused by "intensive penetration by gases from the not yet wholly solidified intrusions," an opinion expressed by Stelzner and Bergeat. He nevertheless considers the process distinctly earlier than the veins and independent of them. The alteration continues to the greatest depths attained. "Kaolinization" near the veins is a wholly

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[1] Bela von Inkey, Nagyag und seine Erzlagerstätten, Buda-Pest, 1885.

Bela von Inkey, De la relation entre l'état propylitique des roches andésitiques et leur filons minéraux, Internat. Geol. Congress, Mexico, 1906.

M. v. Pálfy, Das Goldvorkommen im siebenbürgischen Erzgebirge, etc. *Zeitschr. prakt. Geol.*, 1907, pp. 144–148.

C. Semper, Beiträge zur Kenntniss der Golderzlagerstätten des siebenbürgischen Erzgebirges, *Abh.* K. preuss. geol. Landesanstalt, Neue Folge, Fasc. 33, 1900.

F. Schumacher, Die Golderzlagerstätten der Rudaer Zwölfapostel-Gewerbschaft zu Brád im Siebenbürgen, *Zeitschr. prakt. Geol.*, 1912, pp. 1–85.

different process according to Schumacher and is superimposed upon propylitization. Kaolin and sericite are both present in the altered rock, but the possible influence of descending waters on kaolinization is inadequately treated. The alteration of the wall rock in a vein 0.5 meter thick extends only about 10 centimeters from the vein, but many smaller veins have proportionately wider zones of alteration.

An older set of veins is composed of clastic material of shale and igneous rocks ("Glauch veins"). They are interpreted as having been filled by ascending liquid muds. Similar veins in the Silverton district, Colorado, have been described by Ransome.

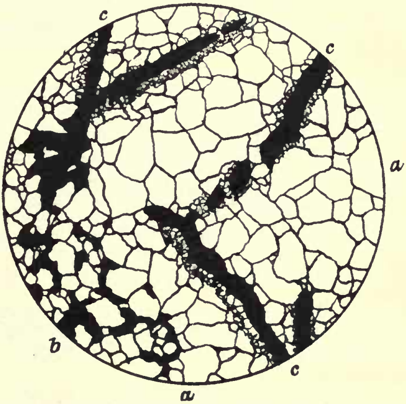

FIG. 155.—Rich gold-bearing quartz, Brád, Transylvania. *a*, Granular quartz; *b*, gold between grains; *c*, plates of gold accompanied by crushed quartz. Magnified. *After F. Schumacher.*

The gangue of the ore-bearing veins is composed of quartz (rarely chalcedony), calcite, rhodochrosite, and barite, a frequently recurring association in veins of this class. The quartz is usually fine-grained, sometimes drusy, cellular, or honeycombed, but not amethysthine. Pyrite in small crystals is abundant; the concentrates contain 10 grams of gold and 69 grams of silver per ton, while the pyrite in the country rock contains 7 to 15 grams of gold and 10 to 15 grams of silver per ton; both kinds are therefore poor in gold. Marcasite has been noted in only one mine, where it occurs on quartz, associated with free gold. Zinc blende is associated with pyrite and is poor in gold but contains more silver than the pyrite. Chalcopyrite and galena where present are poor in gold but contain several hundred grams of silver per ton.

The principal ore mineral is native gold, which occurs commonly in crushed quartz or in little fissures (Fig. 155), or as sheets or wires between the quartz combs of veinlets. Some of it is found in coarse quartz and is apparently older than the quartz or of contemporaneous origin. It occurs also in sheets along the cleavage planes of calcite and in lumps or nodules in clay. In part it is therefore of relatively late origin. The gold contains 28 per cent. silver and the ores average 10 grams of gold per ton. Tellurides and rich silver minerals are rare.

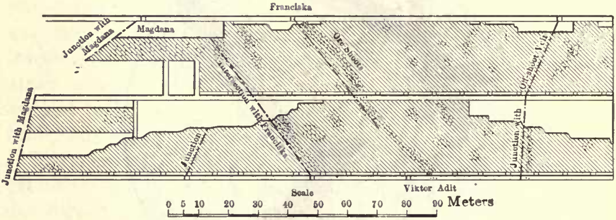

FIG. 156.—Section of stoped area in vein at Brád, Transylvania, showing pockets of rich ore; also rich shoots following intersections with barren veins. *After F. Schumacher.*

The structure of the veins is irregularly massive, though in places crusted, banded, and drusy. Brecciated structures are common. Small prismatic and rectangular pseudomorphs of quartz are considered as replacements of gypsum but strongly resemble the similar casts of celestite at Cripple Creek, Colorado.

The ore-shoots are irregular; sometimes they are narrow but extend with steep dip for 100 or 200 feet vertically. Shoots often occur at junctions and intersections (Fig. 156). At a junction of two veins with a narrow pyritic seam was found a pocket from which in one day gold weighing 55 kilograms was extracted. Near the surface the veins were poor. The richest zone extended from about 100 meters below the surface down to a depth of 320 meters. The remarkable dependence of the rich ore on narrow seams of pyrite is evident and recalls analogous conditions in the Thames district, New Zealand, and the "indicators" of Victorian quartz mines and many other gold deposits (Fig. 157).

B. von Inkey held that the gold was concentrated by leaching from the country rock. Schumacher believed that it was deposited by ascending hot waters in the vicinity of necks of intrusive rocks. Beyond the intrusive necks the veins persist but contain only gangue minerals. M. Dittrich examined fresh and propylitic andesite, using the cyanide process, but found gold in neither.

While much of the gold is distinctly later than the gangue it is difficult to say whether we have to deal here with the effect of descending waters or with the last phases of vein formation. A similar problem is offered in the rich pockets of veins at Thames, in New Zealand; in neither place is enrichment by descending waters satisfactorily proved.

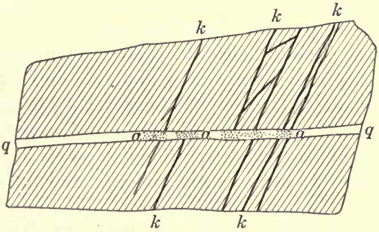

FIG. 157.—Pockets of native gold (*a*) in quartz vein (*q*) along intersections with pyritic seams (*k*). *After F. Schumacher.*

Hauraki Peninsula, New Zealand.[1]—The Hauraki region in the northern island of New Zealand is richly mineralized in several districts. Its rocks consist mainly of andesite and dacite flows of Eocene or Miocene age covered by Pliocene rhyolites.

A production of about $30,000,000 is recorded from the Thames district, though but little gold is now obtained there. The

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[1] James Park, Geology of Hauraki gold field, *Trans.*, N. Z. Inst. Min. Eng., vol. 1, 1897, p. 3.

P. G. Morgan, Geology, etc., of Waihi, *Trans.*, Austr. Inst. Min. Eng., vol. 8, 1902, p. 166.

J. M. Bell and C. Fraser, The great Waihi gold mine, *Bull.* 15, New Zealand Geol. Survey.

A. M. Finlayson, *Econ. Geol.*, vol. 4, 1909, pp. 632–645 (with literature).

Arthur Jarman, The geology of the Waihi-Grand Junction mine, *Trans.*, Inst. Min. and Met. (London), vol. 25, 1916, pp. 3–40, with discussion.

veins are contained in broad belts of soft, propylitic andesite (see p. 480) and dip 40° or more. Great masses of low-grade quartz occur, but the gold is derived mainly from rich pockets occurring down to a depth of 400 to 600 feet below the surface. One of these pockets in the Caledonia mine, about 1871, yielded 9 tons of gold in 15 months. The veins have been followed from a height of 1,500 feet above the sea to 640 feet below it, but owing to intervening faults the real vertical extent is only 1,200 feet. Park states that the veins do not continue into the underlying Jurassic shale and that they are thus limited to the thickness of the lava flows in which they occur. The rich shoots occur mainly where the veins are intersected by small stringers or "leaders." Opinions differ widely as to whether this concentration in pockets is due to descending waters or not. In all probability, however, it was one of the latest phases of the primary mineralization. The principal ore mineral is gold alloyed with 30 to 40 per cent. silver, but some pyrite, chalcopyrite, zinc blende, galena, stibnite, and pyrargyrite also occur. Arsenopyrite and native arsenic, the latter secondary, occur at Coromandel. The gangue minerals, besides quartz, are dolomite and, occasionally, rhodonite.

The Karangahake deposits, 40 miles south of the Thames district, are also in propylitized andesite and dacite but differ somewhat from the type described and consist in brief of calcite and quartz with more or less sulphides. The best known deposits are at Waihi. The Waihi lodes are conspicuous and were discovered in 1878; in part the croppings are covered by rhyolite and the development of the deposit therefore falls between the two eruptions. The ore proved difficult to amalgamate and the mines achieved importance only after the introduction of the cyanide process. To the end of 1917 the total production amounted to about $57,000,000. In 1917 the ore averaged $8 in gold and 1 ounce of silver per ton. The country rock is a green propylitic dacite with some pyrite, calcite, and seams of quartz and adularia. This rock often adjoins the veins without further alteration, but transitions to the quartz filling by silicification are said to exist.

The vein system is complex, and sixteen steeply dipping and interconnecting veins are known. Of most importance is the Martha lode (Fig. 158), a wide fissure vein with brecciated walls; the quartz is formed largely by filling, in part by silicification. On the 500-foot level the lode is in some places 80 feet wide; for half of this width it is barren, but the other half is said to average \$15 to \$20 per ton. The proportion of gold to silver by weight is 1:3 or 1:4 and this average was maintained from the surface down. The water level stood within 200 or 300 feet of the surface. The lode is said to contain ore for a horizontal distance of 1,700 feet. The developments in the deepest levels are said to be disappointing as to the quantity of ore, but the lode itself maintains its strength.

A lively controversy has lately developed in regard to near-surface intrusions. Bell and Frazer consider the dacite intrusive in andesite flows, the inference being that ore may only be expected in that rock. Jarman believes there are no intrusives but only a series of flows.

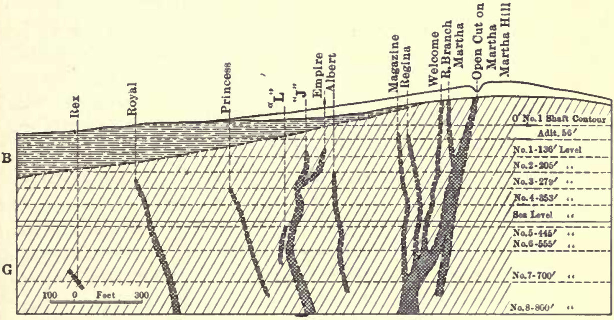

FIG. 158.—Cross-section of Waihi mine, New Zealand, showing lode system in andesite and dacite (*G*), covered by post-mineral rhyolite (*B*). *After C. Fraser.*

A little pyrite was found in the first level in the Martha lode; on the second level the sulphide ore on the foot-wall was a few feet wide; on the 500-foot level 20 feet of sulphide ore was exposed on the foot-wall, while the remainder of the vein, at this place 40 feet wide, was thoroughly oxidized, with much black manganese oxide. This sulphide ore is of nearly the same value as the oxidized ore, containing perhaps a little more gold and a little less silver.

The ore consists of quartz and calcite, with 3 per cent. of pyrite, zinc blende, galena, and argentite. The sulphides are often banded and the gold values are mainly in the pyrite; the bullion contains some selenium. Throughout the oxidized zone the calcite is dissolved, leaving a lamellar quartz ore stained by manganese, but this change is produced mainly by descending waters. In other mines of the district there are indications of a pseudomorphic deposition of silica, similar to that of the De Lamar mine, Idaho (p. 513), by a late phase of ascending solutions. In at least some mines in the Karangahake district the ore becomes poor when the zone of the calcite, unchanged by descending waters, is reached.

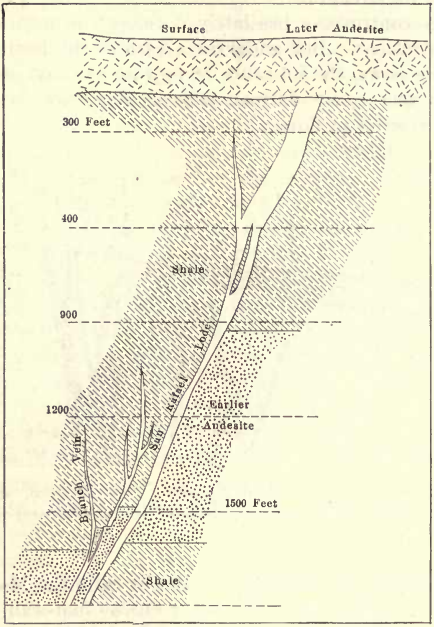

Fig. 159.—Cross-section of the San Rafael lode, El Oro, Mexico, showing branch veins.

The depth of the oxidation in the Waihi mine below water level is noteworthy and probably indicates a dry, intervolcanic epoch.

El Oro, Mexico.—As there are few important gold deposits in Mexico, the occurrence at El Oro, 70 miles northwest of the federal capital, is of special interest. The district is situated on the volcanic high plateau at an elevation of about 10,000 feet. The barren and unaltered andesites of this plateau overlie the ore-bearing formation which consists of a thick flat dipping series of well stratified black bituminous shale with some sandstone; in places these Jurassic sediments are covered by earlier andesites, which near the vein contain pyrite and chlorite and in other places they are intruded by sills of similar andesitic rocks. These earlier andesites are held to be of Miocene age.

The lodes, about ten in number, only outcrop at one or two places and have been opened by cross cut tunnels and shafts. Almost all of the important work has been done since 1904. There are two principal master lodes, the San Rafael and Dos Estrellas striking N.N.W. and dipping steeply S.S.W. The production from the San Rafael alone since 1904 is approximately $40,000,000 from not less than 5,000,000 metric tons of ore. The Dos Estrellas lode for many years yielded about $5,000,000 per annum.

The lodes occupy fault fissures, which in their upper parts at least were open and much of the ore has been deposited by filling of open space. In depth and especially in andesite rock much of the quartz is formed by replacement. In 1913, the greatest depth attained was 2,000 feet; 200 feet of which was in barren cap andesite. The filling consists of fine-grained quartz intergrown with much coarse-grained calcite. There is only a small percentage of pyrite and zinc blende. The gold is never visible and even close panning often fails to bring a color from rich ore.

In the upper levels some stopes are from 60 to 100 feet wide but in depth the lode contracts to smaller dimensions of 3 to 15 feet.

The "branch veins" are an interesting feature of the large lodes at El Oro. They are steep and persistent stringers caused by the settling of the hanging wall (Fig. 159) and are usually rich containing from $15 to $35 in gold and from 5 to 20 ounces of silver to the ton. The ore from the main lode contains only $5 to $15 in gold with 2 or 3 ounces of silver per ton.

The ore shoots in the main lodes are of primary origin and extend horizontally, to a depth of 500 to 700 feet below the capping andesite. The San Rafael has been stoped continuously for one and one-half miles through three properties. These horizontal shoots are probably caused by an upper, impermeable barrier, now eroded, of andesite or clayey rock.

The ore of the branch veins is usually well banded by crustification (Fig. 160) and is much richer in sulphides, mainly zinc blende and pyrite, than the main lode.

Oxidation of the main lode antedates the capping of younger andesite and is practically complete to a depth of 800 feet. The calcite is dissolved and the quartz remains as a porous, cellular mass. The gold being in the quartz there is a considerable apparent enrichment of gold in the oxidized zone. Some silver has probably been leached, but no evidence was found of transportation of gold. There is little manganese in the ore.

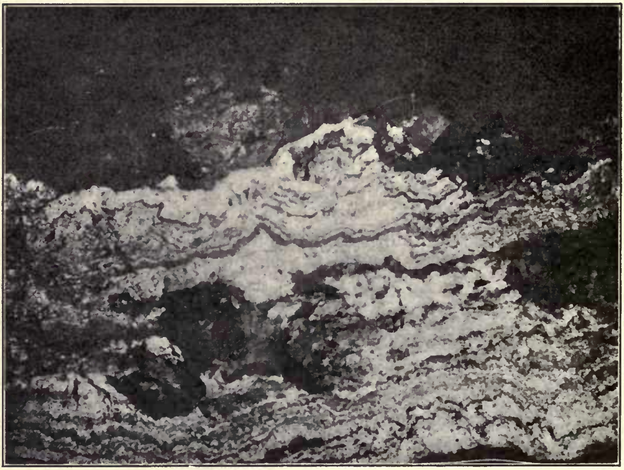

FIG. 160.—Photograph of branch vein on 1,000-foot level, El Oro Mining and Railway Company, showing pronounced banding by sulphides; vein three feet wide contains 9½ ounces gold and 165 ounces silver per ton. Open cavity in middle.

GOLD-QUARTZ VEINS IN RHYOLITE

The Tertiary rhyolites in the Cordilleran region often contain gold-bearing veins These veins are poor in ore minerals other than gold but usually contain some argentite, pyrargyrite, and pyrite. The gold is frequently coarse and accompanied by more or less silver. Among the gangue minerals quartz prevails, but in most cases it is associated with much adularia, probably derived from the surrounding potassic rock. Calcite and fluorite are also common, but barite is rare. The veins are almost characteristically pseudomorphic, with bladed and cellular quartz and adularia, which replace calcite and fluorite. Both veins and sheeted zones occur; in the latter there is little gangue and the gold, as in the Jumbo mine at Hart, California, may be embedded in apparently fresh rhyolite.

There is no real propylitic alteration of the country rock but often extensive silicification and much finely disseminated pyrite. The silicification is attended by concentration of potassium as adularia or sericite. The decomposed upper zone of the veins contains clay seams that may be extremely rich in gold and secondary silver minerals, as at De Lamar, Idaho, and Rawhide, Nevada.

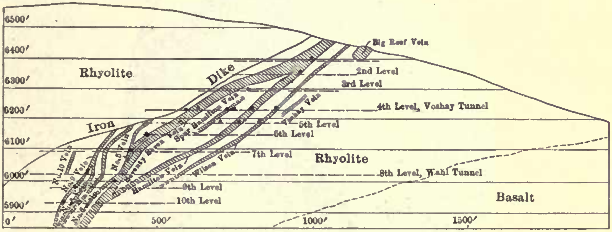

FIG. 161.—Vertical section of the vein system at De Lamar, Idaho.

At the De Lamar mine[1] a series of parallel, gently dipping veins of the kind described abut against a fissure filled with tough clay (the "iron dike") near which the best ore is found (Fig. 161). Below a vertical depth of about 800 feet the values are low, although the veins persist. Free gold is rarely seen. The proportion of gold to silver by weight is 1 :20. The veins are ordinarily 1 to 6 feet thick and distinctly filled, though transitions by silicification were also noted. The filling is wholly quartz, pseudomorphic after calcite, and forms a cellular mass of thin plates coveerd by minute crystals (Figs. 147 and 148). The

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[1] W. Lindgren, *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, p. 122.

value of the ore ordinarily ranges from $10 to $20 per ton. The De Lamar mine has yielded gold and silver to the value of about $7,000,000, but is now closed.

F. C. Schrader has described similar deposits in the Black Mountains[1] in Mohave County, Arizona, and the Jarbidge district,[2] Nevada. At both places adularia is extremely abundant and often forms over 50 per cent. of the gangue (Fig. 162). At the Gold Road mine, in the Black Mountain district, the vein is wide and long; the replacement of calcite and fluorite by quartz and adularia is very clearly shown here. The ore averages $10 per ton and the mine has yielded gold to the value of several million dollars. At Jarbidge a great number of veins have been found intersecting an older rhyolite. An interesting feature is the injection of rhyolite into some of the veins, which are distinctly earlier than the late Tertiary rhyolite flows and were thus formed during a short epoch between two eruptions. Entirely similar veins are found at Rawhide, Gold Circle,[3] Round Mountain,[4] and many other places in Nevada.[5]

FIG. 162.—Thin section showing association of fine-grained quartz (*Q*), with admixed adularia, argentite (*A*), and native gold (*G*), Jarbidge district, Nevada. Magnified 105 diameters. *After F. C. Schrader.*

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[1] *Bull.* 397, U. S. Geol. Survey, 1909.

[2] *Idem,* 497, 1912.

[3] A. F. Rogers, *Econ. Geol.,* vol. 6, 1911, p. 790.
W. H. Emmons, *Bull.* 408, U. S. Geol. Survey, 1910.

[4] F. L. Ransome, *Bull.* 380, U. S. Geol. Survey, 1909, pp. 44–47.

[5] S. H. Ball, *Bull.* 308, U. S. Geol. Survey, 1907, p. 46.

In the Bullfrog district,[1] Nevada, a thick complex of tilted and faulted Miocene rhyolite flows is cut by gold-bearing veins. These veins show various gradations from sheeted zones, in which parallel banded veinlets of alternating crusts of calcite and quartz are separated by thin slabs of rhyolite, through irregular stringer lodes, to lodes made up largely of angular fragments of rhyolite cemented by quartz and calcite. The calcite is in part replaced by cellular quartz, but the process has not been carried to completion.

The extremely rich ore of the National[2] vein, in northwestern Nevada (Fig. 163), has yielded about $3,000,000 from a narrow shoot followed to a depth of 800 feet on the vein. The veins of that district, except for this occurrence, are of the stibnite type. The native gold contains 50 per cent. silver and is more properly called electrum.

FIG. 163.—Dendritic gold (black) in extremely fine-grained quartz of probable colloid deposition. White area is coarsely crystalline comb-quartz, National, Nev. Magnified 15 diameters.

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[1] F. L. Ransome, W. H. Emmons, and G. H. Garrey, *Bull.* 407, U. S. Geol. Survey, 1910.

[2] W. Lindgren, *Bull.* 601, U. S. Geol. Survey, 1915.

ARGENTITE-GOLD-QUARTZ VEINS

Tonopah, Nevada.[1]—The Tonopah district, discovered in 1900, is situated in a group of desert hills in western Nevada about 30 miles north of Goldfield. It is now the most important of the western silver- and gold-producing localities. In 1916 the production amounted to nearly $2,000,000 in gold and 8,700,000 ounces of silver. The ore, which is treated by concentration and cyaniding, yielded $17 a ton. The total output amounts to $92,400,000. In 1917 the production was somewhat smaller.

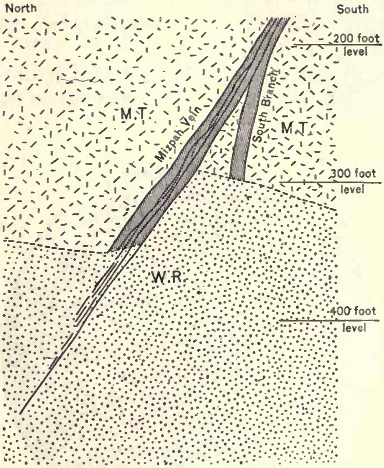

FIG. 164.—Vertical cross-section showing Mizpah vein of first period in Mizpah Trachyte (M. T.), cut off by intrusive sheet of West End rhyolite (W. R.). Later normal faulting has dislocated vein and along the fault fractures mineralization of the second period has taken place prolonging the vein for short distance in the rhyolite. *After J. E. Spurr.*

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[1] J. E. Spurr, Geology of the Tonopah mining district, *Prof. Paper*, 42, U. S. Geol. Survey, 1905, also, *Min. and Sci. Press*, Apr. 22, 1911.

J. A. Burgess, Geology of the producing part of the Tonopah district, *Econ. Geol.*, vol. 4, 1909, pp. 681–712.

A. Locke, The geology of the Tonopah mining district, *Trans.* Am. Inst. Min. Eng., vol. 43, 1913, pp. 157–166.

V. C. Heikes, in *Min. Res.*, U. S. Geol. Survey, annual publication.

J. E. Spurr, Geology and ore deposition at Tonopah, Nevada, *Econ. Geol.*, vol. 10, 1915, pp. 713–769.

According to Spurr the veins intersect a complex volcanic series of flows and near-surface intrusions. The oldest rock is a highly altered trachyte flow (Mizpah trachyte) glassy in its lower part. This is the "earlier andesite" of former reports and contains the valuable veins. Andesite and "West end" rhyolite are intruded into this flow, which was later covered by the "later" or "Midway" andesite flow; at a still later epoch there was a series of flows and intrusions of rhyolite. The most important of the intrusives is the "Tonopah" rhyolite which now outcrops north of the district and is also found in the mine workings. The differentiation of flows and intrusions near the surface is most difficult and have led to conflicting views. Burgess considered the complex to consist of a series of flows. A Miocene age is attributed to the volcanic rocks.

The principal quartz veins are later than the enclosing trachyte and earlier than the following intrusives. They are also older than the "Midway" andesite. A second set of veins was formed after the rhyolite intrusion and before the "Midway" andesite. A few of these are productive. A third set of quartz veins was formed after the intrusion of the Tonopah rhyolite and contain a small amount of sulphides of lead, zinc and copper. All these veins formed at shallow depths and the different types are held to represent various stages of temperature; part of the second and all of the third set of veins are believed to have been deposited at higher temperatures corresponding to the rhyolite intrusions.

The faulting is complex (Fig. 33). Important movements took place at every stage and were caused by the volcanic disturbances.

The importance of the volcanic succession is clearly seen. If an apparently underlying rock is really intrusive and later than the vein formation no continuation of the bonanza veins can be expected in it (Fig. 164). The conditions which resulted in the consolidation of large masses of intrusive magma to more or less glassy rocks seem peculiar and perhaps deserve further investigation.

The productive veins of the earliest period show few outcrops at the surface; they have an easterly strike and various northerly dips. The veins are of moderate thickness though some stopes are 30 feet or more in width. Propylitic alteration (p. 479) affects the trachyte and the andesite but next to the veins there is much silicification with sericite and adularia. The principal gangue mineral is white quartz of fine but variable grain, with banded structure and "chalcedonic appearance," containing parallel bands of finely divided sulphides. The veins are in part filled, but in part appear to have been formed by replacement of the country rock. The primary ore contains some black particles of finely divided gold alloyed with much silver. Argentite and polybasite are the principal ore minerals, with small amounts of pyrite, chalcopyrite, galena, and zinc blende. Selenium is present, probably as a silver selenide. Among gangue minerals there are, besides quartz, rhodonite, adularia, and various carbonates. The secondary ores developed by oxidation and sulphide enrichment are described in Chapter 31. Hübnerite (tungstate of manganese), and scheelite (tungstate of calcium), are believed to belong to the second period of vein formation. The relations of the metals in exceptionally rich concentrates are as follows:

Ag........	25.92 per cent.	Sb........	0.92 per cent.
Au........	0.82 per cent.	Fe.........	9.81 per cent.
Pb........	6.21 per cent.	MgO......	1.49 per cent.
Zn........	5.84 per cent.	CaO.......	3.70 per cent.
Cu........	1.32 per cent.	S.........	not determined.
Se.......	2.56 per cent.	CO_2... ...	6.34 per cent.
As........	0.19 per cent.	SiO_2......	15.18 per cent.

The Comstock Lode.[1]—Among other deposits of this type the Comstock lode deserves special mention. Discovered in 1859, it has yielded, to the end of 1911, a total production of $381,400,000 in silver and gold, of which the gold amounted to $153,000,000. The bonanza period fell in the seventies of the last century and although the production since then has declined greatly, yet during the last few years a systematic unwatering of the deep levels has resulted in a noteworthy rise in output. In 1916 the lode yielded $483,000 in gold and 286,000 ounces of silver, the ore having an average value of $10.64 per ton. The Comstock lode, situated near the summit of the Virginia Range, east of the Sierra Nevada, is a fault fissure of great throw (Fig. 165), traceable two and one-half miles and in places several hundred feet wide, the vein matter of the lode spreading in the hanging wall. Great bonanzas of crushed quartz, in part exceedingly rich in silver minerals, were found at intervals along the lode, especially

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[1] G. F. Becker, *Mon.* 3, U. S. Geol. Survey, 1882.
J. A. Reid, *Bull.* 4, California Univ. Dept. Geology, 1905, pp. 177–199.

in chambers or vertical fissures probably produced by normal faulting of the hanging wall. The greatest vertical depth attained below the outcrop is about 3,000 feet. Mining has been greatly hampered by enormous quantities of hot water containing mainly calcium sulphate. None of the great bonanzas were found below a depth of 2,000 feet.

The lode intersects igneous rocks of deep-seated type, showing transitions and variously classified as diorite, diabase, and augite andesite.[1] These are covered by andesite flows of distinctly

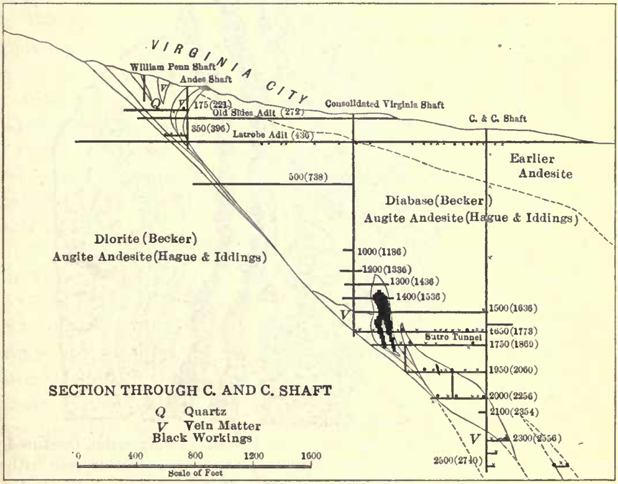

FIG. 165.—Vertical cross-section through the Comstock lode, showing chambered outcrop and bonanza in vertical vein in hanging wall. *After G. F. Becker, U. S. Geol. Survey.*

Tertiary age, which are also mineralized. Both classes of rocks have suffered propylitic alteration, and analyses of the clay gouge near the veins show that sericitic alteration has set in along the principal channels which the solutions followed. The ores consist of quartz and some calcite, in places banded with pyrite, galena, chalcopyrite, zinc blende, and finely distributed rich silver minerals. The valuable minerals are mainly native gold, argentite, stephanite, and polybasite.

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[1] A. Hague and J. P. Iddings, *Bull.* 17, U. S. Geol. Survey, 1885.

There are in places two generations of quartz, as shown in Fig. 166, the older quartz containing principally pyrite. Zeolites are reported in the altered country rock but are apparently not common. According to Reid the descending waters, rich in sulphates, contained notable amounts of gold and silver, and small quantities of these metals were also present in the ascending hot sulphate and carbonate waters. Opinions still differ as to the relative importance of sulphide enrichment in this place.

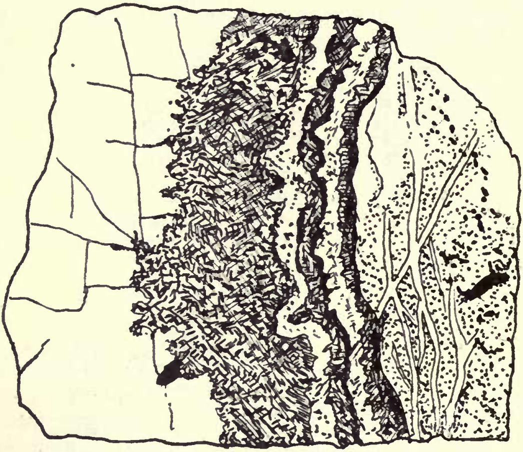

FIG. 166.—Rich ore, Ophir mine, Comstock lode, showing earlier fractured quartz with fine-grained pyrite and some argentite (*a*) and later vein with three generations of galena and argentite (*b*) with some pyrite, chalcopyrite, and quartz (*q*). Drawn from specimen in collection of Massachusetts Institute of Technology, natural scale.

ARGENTITE VEINS

The argentite veins have numerous representatives in Mexico, as at Pachuca, Real del Monte, and Guanajuato. In general they intersect andesitic rocks of supposedly Miocene age but also cut adjacent or underlying Mesozoic sediments.

The rich and long worked veins of Pachuca[1] have come into renewed prominence by the successful application of the

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[1] J. Aguilera and E. Ordóñez, El mineral de Pachuca, *Boletín*, Inst. geol. de México, Nos. 7, 8, 9, 1879.

cyanide process to their ores. A complicated vein system intersects andesite flows covering Cretaceous sediments. The andesite is extensively propylitized and this altered rock also adjoins the veins, near which, however, a silicification is often superimposed upon the chloritization. In places the andesite is entirely silicified. The veins are filled fissures, crustified or brecciated, with quartz, sometimes amethysthine, as the principal constituent of the gangue; there are also rhodochrosite, rhodonite, adularia, and calcite, the last named being the most recent.

The ore minerals consist of argentite, stephanite, polybasite, pyrite, galena, and zinc blende. The veins have been followed to a depth of 2,000 feet and here contain argentite, pyrite, and zinc blende. The ores average about 18 ounces of silver to the ton. The oxidation is marked by the zones of the "colorados" and the "negros," the first of which contains limonite with silver haloid salts and the second much manganese. The "negros" are said to contain more gold than the deeper ores, which are very low in this metal.

There are many other old and famous silver-mining districts in Mexico, the veins of which are similar to those of Pachuca. Among them are Guanajuato, Zacatecas, Sombrerete, Fresnillo, Batopilas, and Parral. In many of these districts, however, the veins are contained in Cretaceous or Jurassic slates and sandstones, though, without much doubt, the mineralization is genetically connected with the surrounding igneous rocks. The latter are in some cases of intrusive origin.

GOLD TELLURIDE VEINS

Cripple Creek.[1]—The veins of Cripple Creek, situated in an otherwise barren part of Colorado, have since 1891 annually yielded a large amount of gold, which in 1900 reached a maximum of $18,000,000. In 1916 the production of gold was valued at $12,120,000, but the silver recovered amounted only to $53,000. The total output of the district to the end of 1916 is $285,-

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[1] W. Cross and R. A. F. Penrose, *Sixteenth Ann. Report*, U. S. Geol. Survey, pt. 2, 1896, pp. 1–209.

W. Lindgren and F. L. Ransome, *Prof. Paper* 54, U. S. Geol. Survey, 1906.

Horace B. Patton, The Cresson Bonanza at Cripple Creek, *Min. and Sci. Press*, Sept. 15, 1917.

500,000. Individual mines have produced from $10,000,000 to $30,000,000 each. The district lies on a granitic plateau a few miles southwest of Pikes Peak, at elevations of 9,000 to 11,000 feet. Within a few square miles are a large number of producing mines; 64 reported in 1916 the production of $46,000 short tons of ore, averaging $12 per ton. In earlier years the average value was $30 to $40 per ton, but during recent years an increasing quantity of low-grade ore from dumps, etc., has been treated, some of it containing only $3 or $4 per ton. Six or eight large mines, among which are the Golden Cycle, the Portland, the Vindicator, the El Paso, and the Elkton, contribute one-half the output, the ore having a value of $18 to $30 per ton. The cyanide process preceded by roasting is now almost universally used for the ores of higher grade; most of the ore is reduced in large mills at Colorado Springs, although there are many smaller cyanide plants in the district.[1]

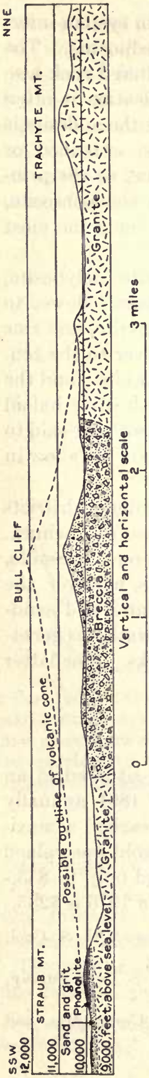

FIG. 167.—Section showing probable outline of the Cripple Creek volcanic cone at the close of the volcanic epoch. *After Lindgren and Ransome, U. S. Geol. Survey.*

The mining operations have always suffered from a large quantity of mine waters and the greatest depth attained is about 2,000 feet. This is in the region of the Portland, Golden Cycle and Vindicator mines. The Roosevelt tunnel, recently completed, is nearly 3 miles long and now drains the mines to an elevation of 8,020 feet—that is, 770 feet below the El Paso drainage tunnel. This tunnel, in 1916, discharged about 10,000 gallons per minute and has now nearly reached the Portland mine at an elevation of 8,112 feet and 164 feet below the 1,900 foot level.

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[1] Data from reports by C. W. Henderson in *Min. Res.*, U. S. Geol. Survey, annual publication.

The rocks constituting the plateau are pre-Cambrian red granite, with some gneiss and fibrolite schist. Breaking through this basement is a mass of Tertiary volcanic rocks, the area having a diameter of 2 or 3 miles. As shown by the mining operations the contact surface between the granite and the volcanic mass is steep, or even vertical or overhanging and there is little doubt that this "plug" of volcanic material represents the core of a Tertiary volcano which formerly rose above the plateau, as tentatively indicated in Fig. 167. The bulk of the remaining core is composed of tuffs and breccias of latite-phonolite and these are cut by dikes and intrusive masses of phonolite and syenite.

FIG. 168.

FIG. 169.

FIG. 168.—Vein filling, Portland mine, Cripple Creek (purple quartz). *f*, Fluorite; *q*, quartz of coarser and finer grain; *p*, pyrite. Magnified 50 diameters.

FIG. 169.—Filled space of dissolution in granite, Independence mine, Cripple Creek (granite ore). *o*, Orthoclase of granite; *m*, biotite replaced by adularia and pyrite; *v*, adularia (secondary orthoclase, showing crustification; *q*, quartz. Magnified 14 diameters.

The latest manifestations of volcanism were basic dikes of monchiquite and vogesite, and the veins were apparently formed soon after these dikes had been intruded. Many of the rocks contain a notable amount of combined water. The deposits are veins which followed a system of roughly radiating, steep fissures (Fig. 37), believed to have resulted from compressive stress developed in a settling volcanic mass. The physiographic history of the district indicates that the surface at the time of vein formation was only from a few hundred feet to 1,000 feet above the present surface.

In the granite adjoining the contact are also found irregular bodies of ore, formed by replacement. Most of the veins are closely spaced sheeted zones (Fig. 42) a few feet wide, though some attain a width of 20 to 40 feet. The ore deposition has generally taken place by filling along the narrow fissures and only to a smaller extent by replacement of the intervening or adjoining rock. Low-grade ores are formed by mineralization of narrow seams in the country rock.

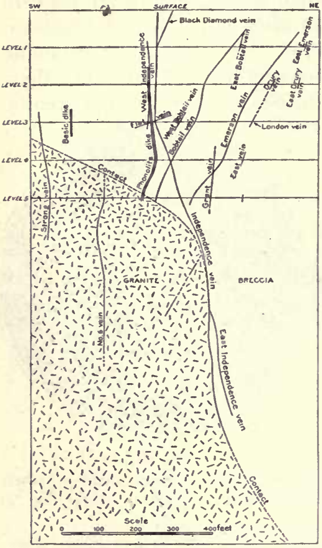

Fig. 170.—Section through Stratton's Independence mine, Cripple Creek, showing relation of veins to granite-breccia contact. *After Lindgren and Ransome, U. S. Geol. Survey.*

There is little or no native gold, except in the oxidized zone (p. 830). The principal ore mineral is calaverite ($AuTe_2$, with but little silver), of yellowish white color and often well crystallized. Associated with this are small quantities of pyrite, zinc blende, tetrahedrite, stibnite, and molybdenite, rarely hübnerite. The gangue consists of quartz and fluorite, with some dolomitic carbonate. The fluorite, quartz, and calaverite are often intergrown, forming a fine-grained rock which has the purple color of the fluorite (Fig. 168). The vein structure is drusy and the calaverite was among the latest minerals formed. The vein filling consists of about 60 per cent. quartz, 20 per cent. dolomite, 20 per cent. fluorite, 0.1 per cent. gold, and 0.2 per cent. tellurium, with iron, copper, zinc, and molybdenum present in fractions of 1 per cent. Extremely rich pockets of calaverite are sometimes found. The most notable instance is the Cresson bonanza referred to in the list of literature (p. 521).

The replacement ore consists of the ordinary red granite, often drusy and partly replaced by adularia, fluorite, and calaverite

(Fig. 169). In the upper levels there are a great number of short veins all of which carry more or less ore. In the aggregate these veins contain an enormous amount of ore, some of which is of exceedingly high grade. The veins are less abundant in the lower levels and some of them are of lower grade. Many rich veins continue, however, to the lowest levels.

As shown in the report cited the ore-shoots are to a marked degree influenced by intersections with other veins or dikes, but many of the largest and richest shoots have no such relation.

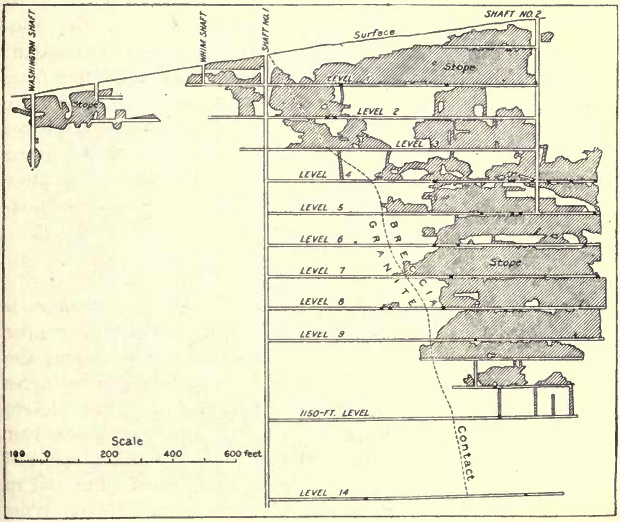

FIG. 171.—General north-south section through Stratton's Independence mine, Cripple Creek, showing stopes on Independence vein. *After Lindgren and Ransome, U. S. Geol. Survey.*

The greatest horizontal extension of a shoot is 1,300 feet. Many shoots terminate in depth, while others have continued to the greatest depths attained. The relations at Stratton's Independence mine are illustrated in Figs. 170 and 171.

The tuffs and breccias are generally altered and contain some fine-grained pyrite, which has little value; the dark silicates alter to carbonates, fluorite, and pyrite and the feldspars to sericite and adularia. Cross and Penrose thought this propylitic alteration earlier than the veins, while Lindgren and Ransome consider it to be caused by the same kinds of solutions that filled the fissures. Similar differences of opinion have been expressed in relation to the propylitization at other places (p. 483). The alteration close to the veins is remarkably slight at Cripple Creek.

There is no evidence that there has ever been an active circulation of surface water in the district. The porous breccias and in general the whole volcanic plug are filled with stagnant water, while there is little water in the surrounding granite. The general conclusion of Lindgren and Ransome is that the vein-forming epoch was brief and that the remarkable and abundant telluride ores were formed by alkaline solutions emanating from deeper igneous masses, the last effects of these emanations being the exhalations of carbon dioxide and nitrogen, which have not yet subsided. The waters ascended rapidly in the deeper parts of the volcanic plug, but near the surface they spread out in more numerous fissures and precipitation followed by cooling or mixture with descending solutions.

GOLD SELENIDE VEINS

Occurrence of Selenides.—In minute quantities selenium is present in many deposits, particularly in the pyritic copper deposits, and it is recovered on a rather large scale during the electrolytic refining of copper. As distinct minerals the selenides are apparently confined to the metallic veins formed at moderate or shallow depths. Their presence in some rare quicksilver deposits has already been mentioned. In the silver veins of Mexico selenides of silver and lead have been found; and in some silver-gold veins like those at Tonopah or gold-silver veins like those at Waihi they are important constituents. At both Tonopah and Waihi other minerals are present in quantities.

The type of veins described in these paragraphs is remarkably free from ore minerals other than native gold and selenides, and it is rare, only two examples being known, that of Republic, in Washington, and that of Radjang Lebong, in Sumatra. In some respects, however, the Tonopah veins are allied to this type.

In both the places mentioned there is a predominating gangue of very fine-grained quartz, beautifully banded by crustification, but not markedly drusy; it has a "chalcedonic" appearance, although there is really but little chalcedony present, and it resembles strongly, both in hand specimen and in thin section, some kinds of sinter deposited at the orifices of hot springs. The gold is present in very fine distribution and the gold selenide has not yet been positively identified. These veins have probably been deposited close to the surface.

Republic, Washington.—At Republic[1] a series of Miocene andesite and latite flows filling an old valley have been intruded by a mass of latite porphyry belonging to the same general period of eruption. Over a considerable area the andesite and porphyry have suffered normal propylitization, chlorite, earthy carbonates, and pyrite being the principal minerals formed. A series of parallel fractures dipping from 38° to 80° have been opened in the volcanic rocks and are occupied by sharply defined veins averaging 3 or 4 feet in width. Against these the propy-

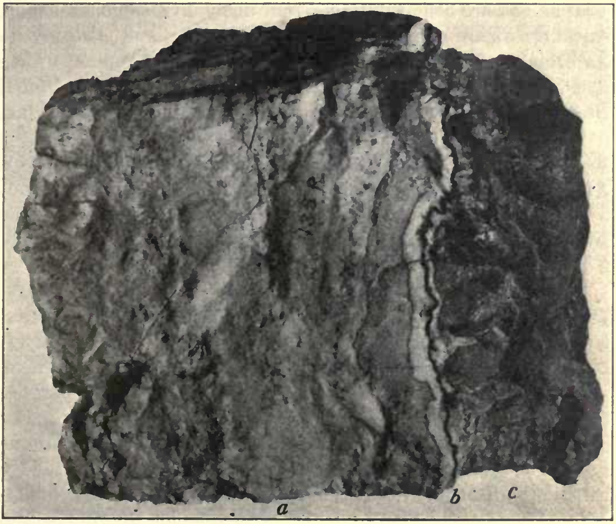

FIG. 172.—Typical ore, Republic mine, Republic, Washington. *a*, Fine-grained quartz, banded; *b*, streak of black, finely divided sulphides and selenides (?); *c*, altered latite porphyry. Natural size.

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[1] Howland Bancroft and W. Lindgren, *Bull.* 550, U. S. Geol. Survey, 1914

litic rock borders, usually with little further alteration. The banded filling (Fig. 172) consists of quartz and calcite and also includes dark masses of jasperoid of uncertain derivation. There is some adularia in the filling and in a few places can be seen the beginning of a replacement of calcite by fine-grained quartz and adularia. In one mine the quartz filling has been replaced by laumontite containing much silver.

Free gold is rarely visible, but the valuable portions of the veins lie along narrow dark bands that are parallel to the crustification and are believed to represent finely divided gold selenide. Local crusts are rich in free gold, tetrahedrite, and chalcopyrite and this material contains about 2 per cent. of selenium, which, according to experiments by Dr. Chase Palmer, of the United States Geological Survey, is probably combined with gold.

The ores of Republic have proved difficult to treat by the cyanide process. Their grade varies greatly, averaging perhaps $11 per ton. The proportion of gold to silver by weight is about 1:3. Oxidation has resulted in small changes but has set some silver free. Since 1897 the district has yielded about $7,000,000. The greatest depth attained is 800 feet.

Sumatra.—The Radjang-Lebong field,[1] in southern Sumatra, has yielded much gold in recent years. The annual production is about 50,000 ounces of gold and 300,000 ounces of silver. Andesite is the country rock and the principal vein, which has a width of 17 feet, is divided into five well-defined seams separated by silicified andesite. According to R. Beck the bluish-gray quartz is beautifully banded in thin concentric crusts of "fibrous quartz."[2] The rich ore, like that of Republic, is indicated by thin dark dendritic crusts similar in appearance to the quicksilver selenide from Lerbach, the silver selenide at Tillingerode (both localities in the Harz Mountains), and the copper selenide of Skrikerum, in Sweden.

The ore, of which only a small part is amenable to amalgamation, contains on an average 41 grams of gold and 318 grams of silver per metric ton. There is a little pyrite and chalcopyrite. The bullion, according to Truscott,[3] contains

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[1] R. Beck, Lehre von den Erzlagerstätten, vol. 1, 1909, p. 488.

[2] The expression "fibrous quartz" used by Beck suggests a possible replacement of primary calcite by quartz and adularia.

[3] M. Maclaren, Gold, London, 1909, p. 298.

S. J. Truscott, *Trans.* Inst. Min. and Met., London, vol. 10, 1902, p. 53.

Gold and silver	91.52
Selenium	4.35
Copper	1.82
Lead	1.65
Zinc	0.48
Iron	0.14
Total	99.96

THE BASE-METAL VEINS

Among the deposits formed relatively near the surface by ascending thermal waters in genetic connection with igneous rocks, ores rich in the baser metals and worked principally for these metals are rather exceptional. Heavy deposits of pyrite and chalcopyrite ore are very seldom found; one instance is furnished by the Nacozari district, in Mexico, where at the Los Pilares mine[1] a large body of low-grade pyrite and chalcopyrite is worked. The ores occur mainly in the interstices of a brecciated rhyolite in a series of fractures arranged in almost circular form. The pyrite is often crystallized; the gangue is quartz; the ore is poor in gold and silver.

The veins and stocks of the San Juan region, Colorado, described in more detail below, are sometimes rich in lead, zinc, and copper, but yield principally gold and silver. Both galena and zinc blende occur, but the copper is derived mainly from tetrahedrite or enargite.

The Schemnitz deposits, in Hungary, consist of a strong vein system intersecting rhyolite and andesite above Triassic slates and Eocene strata. The gangue minerals are fine-grained quartz and amethyst, together with later calcite, ankerite, rhodochrosite, rhodonite, siderite, and barite, with much pyrite, galena, chalcopyrite, and zinc blende. Among the rarer minerals are adularia, fluorite, and diaspore. The proportion of gold to silver by weight is 1:23.

The Bull-Domingo and Bassick deposits,[2] at Silver Cliff, Colorado, yielded principally gold and silver. The ore consisted, however, largely of sulphides and tellurides, which in the Bassick mine occurred in what is considered a volcanic neck. The Bassick deposit was mined to a depth of 800 feet and yielded

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[1] S. F. Emmons, *Econ. Geol.*, vol. 1, 1906, pp. 629–643.

[2] S. F. Emmons, The mines of Custer County, Colorado, *Seventeenth Ann. Rept.*. U. S. Geol. Survey, pt. 2, 1896, pp. 430–447.

rich ore. The cross-section covered about 25 by 100 feet and the ores encrusted the fragments of volcanic rocks filling the shoot.

When the solutions depositing veins in volcanic rocks leave the flows and enter into the surrounding limestones and other sedimentary rocks, deposition by selective precipitation comes into play and ores rich in sulphides, particularly galena, may be formed. Examples of this are found in several deposits of the Ouray district, Colorado, described by J. D. Irving.

THE SAN JUAN REGION, COLORADO[1]

General Features.—One of the most interesting metallogenetic provinces is that of the rugged San Juan region, in southwest

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[1] W. Cross and C. W. Purington, *U. S. Folio* 57, Telluride; Cross, Spencer, and Purington, *U. S. Folio* 60, La Plata; Cross, Howe, and Ransome, *U. S. Folio* 120, Silverton; Cross, Spencer, and Ransome, *U. S. Folio* 130, Rico; Cross, Howe, Irving, and W. H. Emmons, *U. S. Folio* 131, Needle Mountains; Cross and Hole, *U. S. Folio* 171, Engineer Mountain; Cross and Howe, *U. S. Folio* 153, Ouray.

C. W. Purington, Report on mining industries, Telluride quadrangle, *Eighteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1897, pp. 745–850.

Cross and Spencer, Geology of the Rico Mountains, *Twenty-first Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1900, pp. 7–165.

F. L. Ransome, Economic geology of the Silverton quadrangle, *Bull.* 182, U. S. Geol. Survey, 1901.

J. D. Irving, Ore deposits of the Ouray district, *Bull.* 260, U. S. Geol. Survey, 1905.

W. H. Emmons, Neglected mine, etc., Durango quadrangle, *Bull.* 260, U. S. Geol. Survey, 1905.

J. D. Irving and H. Bancroft, Geology and ore deposits near Lake City, *Bull.* 478, U. S. Geol. Survey, 1911.

Arthur Lakes, Geology of western ore deposits, Denver, 1905.

T. A. Rickard, Across the San Juan Mountains, New York, 1903.

T. A. Rickard, The Enterprise mine, Rico, Colo., *Trans.* Am. Inst. Min. Eng., vol. 20, 1897, pp. 906–980.

C. W. Purington, The Camp Bird mine, *Trans.* Am. Inst. Min. Eng., vol. 33, 1904, pp. 499–528.

C. W. Purington, Ore horizons in the San Juan Mountains, *Econ. Geol.*, vol. 1, 1905, pp. 129–133.

A. Winslow, The Liberty Bell mine, *Trans.* Am. Inst. Min. Eng., vol. 29, 1900, pp. 285–307.

J. E. Spurr, summary in *Prof. Paper* 63, U. S. Geol. Survey, 1908, pp. 111–168.

W. H. Emmons, A preliminary report on the geology and ore deposits of Creede, Colo., *Bull.* 530, U. S. Geol. Survey, 1912, p. 42–65.

C. W. Henderson in *Min. Res.*, U. S. Geol. Survey, annual publication.

Colorado, including the mining districts of Telluride, Ouray, Silverton, Lake City, Rico, Needle Mountains, La Plata, and Creede (Fig. 173). Space is not available to describe these fully, but some of the more important relations should be pointed

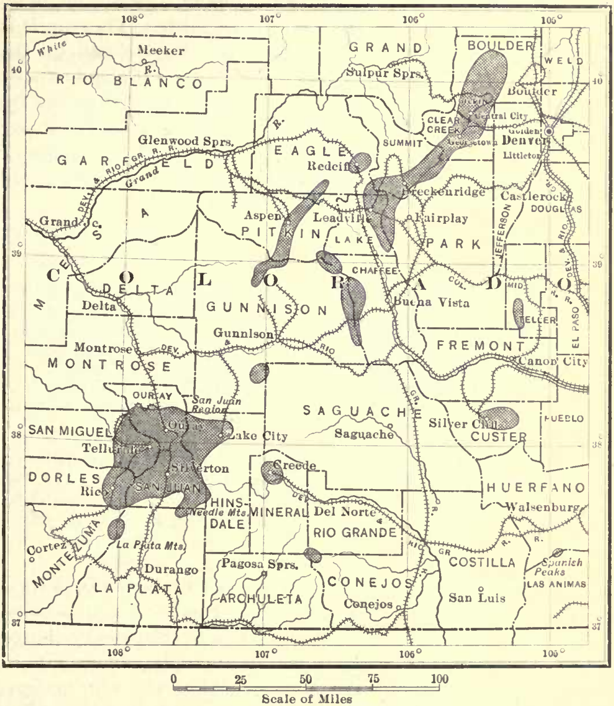

FIG. 173.—Map showing approximate distribution of the principal mining regions in Colorado. *After J. E. Spurr, U. S. Geol. Survey.*

out. The districts are known mainly through the labor of the geologists of the Federal Survey.

The San Juan Mountains consist chiefly of volcanic rocks, poured out over a basement of Paleozoic and Mesozoic sediments; pre-Cambrian rocks are also exposed in places (Fig. 174). The volcanic flows, occupying over 3,000 square miles, have a total thickness of many thousand feet and were erupted at intervals during the whole Tertiary period. The lowest formation, called the San Juan tuff, consists of about 3,000 feet of andesitic material. Covering this are andesites, latites, and rhyolites, called the Silverton series, 4,000 feet thick; this in turn is overlain by the Potosi series, consisting of rhyolite and latite. The eruptions were separated by epochs of erosion and were closed by the effusion of the Hinsdale series of rhyolites, andesites, and basalts.

Deep erosion has laid bare these flows to a depth of several thousand feet and exposed a number of smaller intrusive stocks and sheets of dioritic or monzonitic character and granular or porphyritic texture. The very latest intrusions are represented by dark dikes of lamprophyric character.

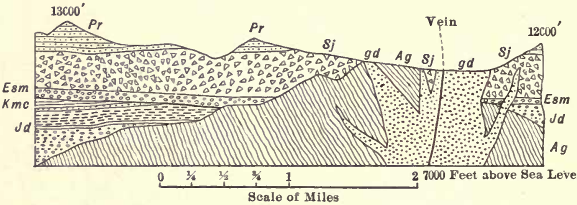

FIG. 174.—Section at Telluride, Colorado. *Ag,* Algonkian quartzite; *Jd,* Dolores formation (Jurassic); *Kmc,* Mancos shale (Cretaceous); *Esm,* San Miguel conglomerate (Eocene); *Sj,* San Juan tuff; *Pr,* Potosi rhyolite; *gd,* gabbro-diorite; *Sj, Pr, gd,* Eocene and Miocene. *After W. Cross, U. S. Geol. Survey.*

This vast uplifted and eroded dome of volcanic rocks is traversed by numerous systems of strong fissure veins, many of them continuous for several miles. Their development followed closely after the latest epoch of volcanic activity and they intersect several of the intrusive masses. While the vein systems bear the marks of deposition within a moderate distance of the original surface, there are some features that tend to connect them with deposits formed at greater depth and in more direct genetic connection with igneous intrusions. Many types are represented, including normal veins, stocks, replacement deposits in sedimentary formations, and some small contact-metamorphic deposits near the contacts of the intrusive masses. Some of the veins are exposed over vertical distances of several thousand feet. The Virginius vein, above Ouray, for instance, has been mined for a vertical distance of 3,300 feet; the Revenue tunnel is 2,400 feet below the outcrops.

The predominating values are in gold and silver, but in some districts much lead, copper and zinc are also present, mainly as galena, tetrahedrite, and zinc blende. The gangue is mainly quartz, but rhodochrosite, rhodonite, barite, and fluorite are often present.

Telluride District.—In the Telluride district the present annual production is about $2,000,000 in gold and 1,000,000 ounces of silver, with some copper and lead, mainly from the strong veins worked by the Liberty Bell, Smuggler-Union, and Tomboy mines. The veins are filled fissures, averaging 3 or 4 feet in width, with crustification banding, drusy structure, sericitization, and silicification of the walls. The larger lodes appear often as a number of parallel plates of filled veins separated by sheets of altered rock. The Smuggler vein, for instance, cuts through the San Juan tuff, the andesite, and the rhyolite, a vertical distance of 2,000 feet (Fig. 175). What aspect the veins assume in the underlying sedimentary formation is as yet unknown.

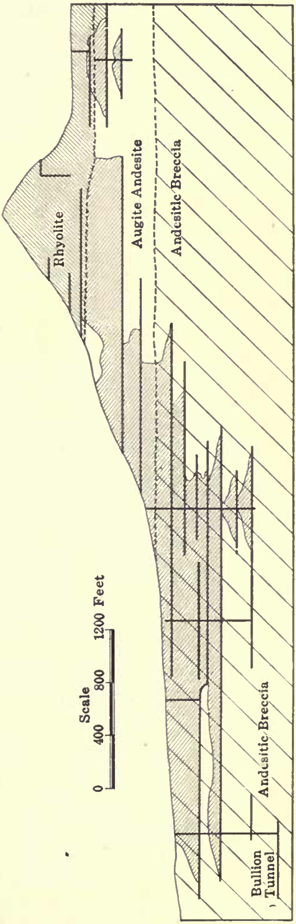

FIG. 175.—Longitudinal section of the Smuggler vein, Telluride, Colorado, showing stopes. After C. W. Purington, U. S. Geol. Survey.

The gangue minerals are quartz, calcite, siderite, rhodochrosite, adularia, barite, and fluorite, the last being abundant in the Tomboy mine (Fig. 176). Native gold, pyrite, galena, zinc blende, and chalcopyrite are the principal metallic minerals. The ores contain 2 or 3 per cent. sulphides and yield about $6 in gold and a few ounces of silver per ton. The treatment consists of a combination of amalgamation, concentration, and cyaniding. There are several minor silver veins and replacement deposits in the district.

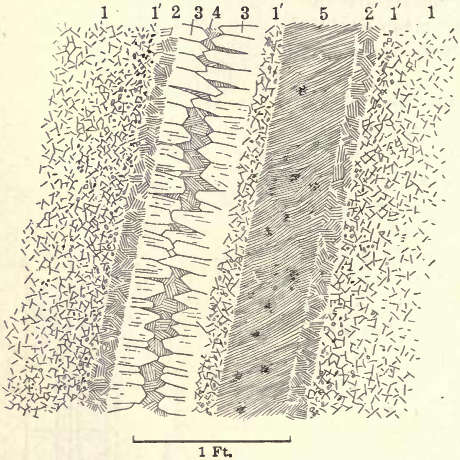

Fig. 176.—Section showing succession of ore minerals in Mendota workings, Smuggler vein Telluride, Colorado. 1, Country rock; 1′, sericitized and impregnated country rock; 2, zinc blende with calcite; 2′, zinc blende with galena; 3, white quartz; 4, rhodochrosite; 5, blue quartz with finely disseminated sulphides. *After C. W. Purington, U. S. Geol. Survey.*

Silverton District.—The Silverton district is rich in veins, which contain more base metals than those of the Telluride district. The present annual production is about $500,000 in gold, 500,000 ounces of silver, 3,500,000 tons of lead, and some copper and zinc. The deposits are simple veins or lodes, averaging about 3 feet in thickness. The structure is commonly massive but sometimes banded (Fig. 177) and drusy; on the whole it

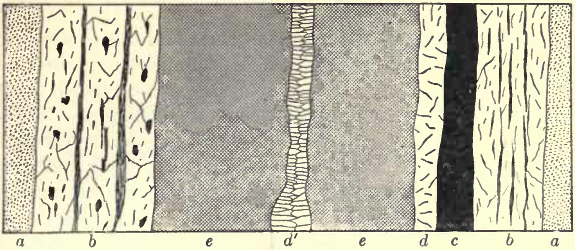

FIG. 177.—Cross section of banded vein near London shaft, Silverton, Colorado. *a*, Country rock; *b*, quartz and chalcopyrite; *c*, tetrahedrite; *d*, *d'*, quartz; *e*, galena. Vein 6 inches wide. *After F. L. Ransome, U. S. Geol. Survey.*

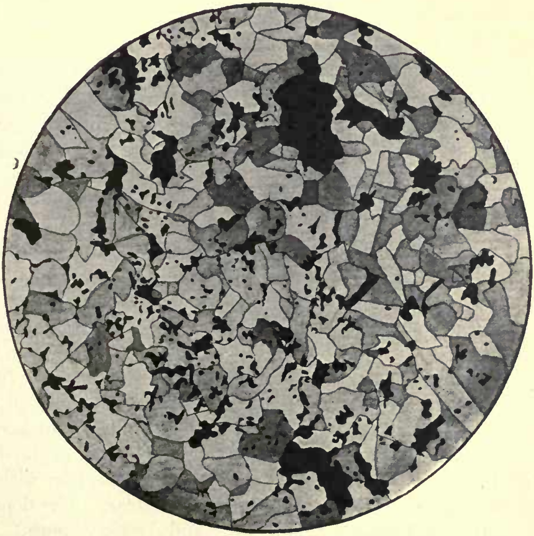

FIG. 178.—Thin section of ore from Ridgeway mine, Silverton, Colorado. Large black areas, pyrite; small black areas, argentite, with a little galena and zinc blende; shaded grains, quartz. Magnified 17 diameters. *After F. L. Ransome, U. S. Geol. Survey.*

resembles that of deeper-seated veins. Many of the veins are rich in sulphides.

The prevailing gangue mineral is quartz and this is of coarser-grained texture than is common in the veins deposited near the surface (Fig. 178). The gangue also includes much calcite, dolomite, rhodochrosite, rhodonite, barite, and fluorite. The ore minerals are pyrite, galena, chalcopyrite, zinc blende, tetrahedrite, enargite, argentite, and native gold, more rarely hübnerite, molybdenite, and various sulphantimonides and bismuthides. Tetrahedrite and galena are very abundant. Tellurides are rare. Rock alteration in this district is discussed in some detail on pages 486–487.

Ouray District.—The Camp Bird lode, in Ouray County, probably represents the continuation of one of the Telluride lodes. For many years the Camp Bird mine has yielded rich returns, amounting from 1903 to 1910 to $16,500,000, principally in gold, though some silver, copper and lead are recovered. The ore contains about $22 in gold per ton. The lode intersects San Juan tuff and andesite and is described as a sheeted zone 4 or 5 feet thick made up of alternating fissure filling and altered rock; filling was, however, the predominating process. The gangue is quartz, often crusted and banded, with rhodochrosite, calcite, and fluorite. The metallic minerals are very fine native gold, with a few per cent. of galena, pyrite, and zinc blende, also some finely distributed tellurides. Mining was discontinued in 1916.

Irving describes a number of replacement deposits in the Ouray district just outside of the volcanic area. They are contained in quartzite or limestone below impervious beds and yield sulphides, ores with galena, tetrahedrite, chalcopyrite, jasperoid, and barite.

Rico District.—The Rico Mountains are a domelike uplift of sedimentary rocks ranging from Algonkian to Jurassic in age, intruded by stocks, sheets, and sills of monzonite, or monzonite porphyry (Fig. 179). The ores are therefore rather of the deep-seated type, but nevertheless their genetic connection with the other deposits of the San Juan region is clear. The deposits form lodes, bed-veins (blankets), and replacements. The blankets often lie parallel to the sheets of intruded rocks or below impervious shales. The abundant ore minerals consist of pyrite, galena, zinc blende, and tetrahedrite, in a gangue of quartz,

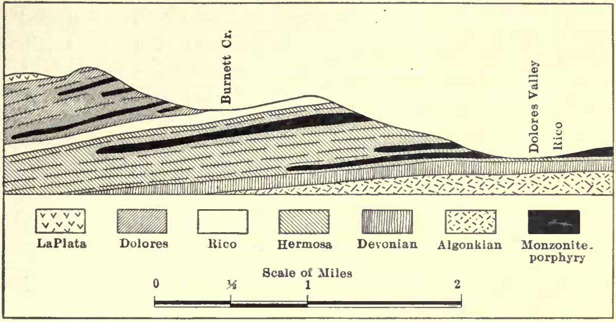

Fig. 179.—Geological section through part of the Rico dome, Colorado. *After Cross and Spencer, U. S. Geol. Survey.*

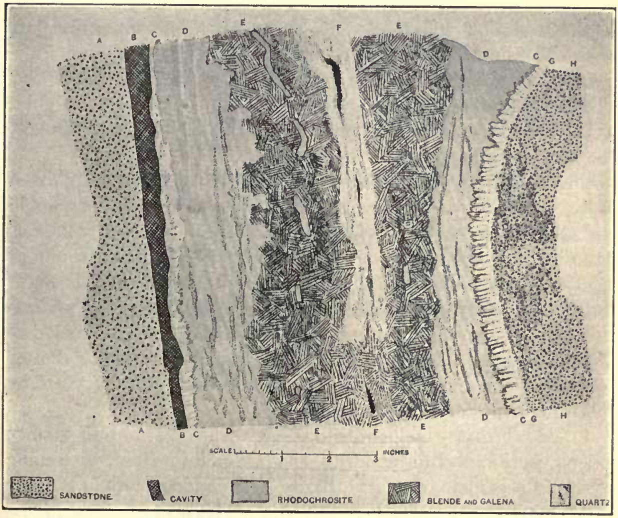

Fig. 180.—Banded ore, Rico, Colorado. *After T. A. Rickard.*

rhodochrosite, calcite, and fluorite. The filling is often beautifully banded (Fig. 180). A very limited vertical range is characteristic of the deposits. The greater part of the production has come from the blankets, a short distance below which the veins have become impoverished in the Hermosa (Pennsylvanian) formation of sandstone, shale, and limetone. Besides the silver-lead ores the district contains larger bodies of low-grade pyritic ores. The rich silver minerals were mainly the result of oxidation and sulphide enrichment of the galena and tetrahedrite.

La Plata, Durango, and Needle Mountains Quadrangles.—South of Rico the sandstones of the Jurassic and Triassic (Dolores formation) are abundantly intruded by sheets of monzonitic rocks. Late dikes of basic character intersect the monzonite porphyries. Veins cutting both porphyries and sediments contain native gold, tellurides (petzite, sylvanite, and calaverite), argentiferous tetrahedrite, tennantite, stephanite, amalgam, pyrite, marcasite, chalcopyrite, galena, and zinc blende. In the gangue quartz prevails, with some rhodochrosite and barite The similarity to the telluride veins of Lake City is striking. Entirely similar telluride veins occur in the northeastern part of the Needle Mountains quadrangle south of Silverton.

Lake City District.—The production of this district is now small, but it has yielded a total output of $1,250,000 in gold, 4,000,000 ounces of silver, and 40,000 tons of lead, the value of the lead exceeding that of the silver.

The deposits range from simple veins to more complicated lodes, but filling with banding was the predominating process in their development. The width ranges from a few inches to 20 feet. Many of the fissures are short and were found to pinch out at relatively slight depth. The alteration of the wall rock is slight but is marked by some silicification and sericitization. Pyritization of the surrounding rhyolite or andesite extends to considerable distances. The veins intersect all rocks below the Potosi series and also cut the intrusions of monzonite porphyry. The following types of veins are recognized:

1. Tetrahedrite-rhodochrosite veins. The ores contain mainly galena, pyrite, and tetrahedrite, the last rich in silver, with some pyrite in a gangue of quartz, barite, and rhodochrosite.

2. Quartz-galena-zinc blende veins. The ores of these veins contain dominant galena, pyrite, and zinc blende, with subordinate chalcopyrite, in a quartz gangue.

3. Telluride veins. The few veins of this type contain gold and silver tellurides disseminated through a quartz gangue with subordinate sulphides, tetrahedrite, and barite.

Creede District.—In the Creede district, in the eastern part of the great volcanic area, strong and beautifully banded veins intersect rhyolite. They carry, in a gangue of amethystine quartz, barite, and some fluorite, a considerable amount of galena and zinc blende. The average ore in 1910 contained 0.09 ounce of gold and 12.29 ounces of silver per short ton, 0.02 per cent. copper, 6.5 per cent. lead, and 1.9 per cent. zinc.[1] Some of the ores are considerably richer in galena. Most of the ore is concentrated, and the total production in 1910 had a value of about $1,000,000. In 1916 the total value was $471,000.

Summary.—The deposits of the San Juan region consist of gold-bearing quartz veins, gold-telluride veins, and base-metal veins generally carrying galena and tetrahedrite. The descriptions show a merging of the types and certain minerals common to most of them, such as rhodochrosite, rhodonite, and barite. The ores are usually well banded and crustified.

The ores are clearly independent of the character of the country rock, as has been already noted by J. D. Irving. They also intersect the intrusive monzonites, where these are exposed by erosion, and they are apparently derived from a deeper source. The ores were probably deposited in one main epoch following the intrusive activity.

Spurr believes that fissuring and mineralization were caused by and followed the dome-like uplifts and that these were due to deep-seated intrusions, not yet exposed by erosion. The metals in the deposits would then have been derived from the ascending emanations of these deep-seated intrusives. The region shows, as few others do, the relation of the upper and deeper vein zones.

GOLD-ALUNITE DEPOSITS

General Features.—In volcanic regions it is not uncommon to find considerable areas of bleached and altered lavas which contain more or less alunite ($K_2O.3Al_2O_3.4SO_3.6H_2O$), an earthy or compact, rarely coarsely crystalline mineral of inconspicuous appearance (p. 479). Occasionally diaspore or gibbsite is associated

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[1] C. W. Henderson in *Min. Res.*, U. S. Geol. Survey, pt. 1, 1910, p. 424.

with alunite.[1] In most cases this basic sulphate, which is insoluble in water, is probably formed by the action of waters containing free sulphuric acid on aluminous rocks. It is also found in places in the oxidized zones of veins containing pyrite.[2] In such altered zones in volcanic rocks alunogen, jarosite, halotrichite, and other sulphates of iron and aluminum are often encountered as products of solution and oxidation.[3] Pyrite sometimes appears as a primary constituent, its iron being probably derived from the ferromagnesian silicates of the rock.

Although the alunite itself is used, in large deposits, for the production of alum and similar salts, it is unusual to find rare metals associated with areas of alunitization. Only one such deposit has been discovered, and that is the remarkable bonanza of Goldfield, described by F. L. Ransome.[4]

Goldfield, Nevada.—The Goldfield district, discovered in 1902, lies in a low range of desert hills in western Nevada. The total production to the end of 1917 was $80,700,000 in gold and 1,250,000 ounces of silver. In 1911 the district yielded gold and silver valued at $10,300,000; in 1917 only $2,000,000.

The geological features consist of a succession of volcanic flows, of which 15 members are recognized, resting on a basement of granite and Cambrian shale. The age of these lava flows and intercalated lacustrine beds probably ranges from Eocene to the latest Pliocene. On the basement rest several flows of rhyolite; then come

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[1] Whitman Cross, Geology of Silver Cliff and the Rosita Hills, *Seventeenth Ann. Report*, U. S. Geol. Survey, pt. 2, 1896, pp. 263–403.

L. de Launay, La métallogénie de l'Italie, 10th Int. Geol. Congress, Mexico, 1907, pp. 125–132.

E. S. Larsen, Alunite in the San Cristobal quadrangle, Colorado, *Bull.* 530, U. S. Geol. Survey, 1913, pp. 179–183.

R. T. Hill, Camp Alunite, Nevada, *Eng. and Min. Jour.*, vol. 86, 1908, pp. 1203–1206.

[2] W. Lindgren and F. L. Ransome, *Prof. Paper* 54, U. S. Geol. Survey, 1906, p. 125.

For more complete list of literature see F. L. Ransome, *Prof. Paper* 66, U. S. Geol. Survey, 1909. p. 132, and B. S. Butler and H. S. Gale, Alunite, a newly discovered deposit near Marysvale, Utah, *Bull.* 511, U. S. Geol. Survey, 1912.

[3] C. W. Hayes, *Bull.* 315, U. S. Geol. Survey, 1907, p. 215.

[4] *Prof. Paper* 66, U. S. Geol. Survey, 1909.

Econ. Geol., vol. 5, 1910, pp. 301–311, 438–470.

[5] Alunitized areas have been described from Vancouver island, in Jurassic and Triassic lavas. These altered rocks contain a little gold. See C. H. Clapp, *Econ. Geol.*, vol. 10, 1915, pp. 70–88.

heavy masses of andesite or latite. In the andesite is intruded a thick sheet of dacite in which most of the ore-bodies are found. The andesite is overlain by 1,000 feet of lacustrine beds, and still later than these are thin flows of basalt and rhyolite. The dacite sheet, almost 700 feet thick, consists of a rock of intermediate composition with glassy groundmass. Silica amounts to 60 per cent., and the potash about equals the soda. The pre-lacustrine rocks are considered Eocene. The lacustrine conditions ended by deformation and a low dome-like uplift resulted.

The principal producing area is small; most of the deposits are in the dacite, though a few are contained in rhyolite, andesite, or latite. The deposits are probably younger than the lake beds but older than the latest rhyolite and basalt; the age is considered late Miocene or early Pliocene. They were thus formed in intervolcanic time, like so many others of the ore deposits connected with effusive rocks. The deposits are contained in an area of highly altered rock and are irregular silicified and fractured rock masses to which the term "ledges" is applied; though irregular (Fig. 181), the ledges show a more or less marked elongation in one direction and are often vertical. The ledges occur in dacite, in andesite, and in some of the rhyolites; they are then parts of the country rock which have suffered more silicification than the adjoining rock, which is also altered. The altered rock, below the zone of oxidation, is a porous dark-gray material with softer white spots to indicate the feldspar phenocrysts.

The four minerals characteristic of this type of alteration are quartz, kaolinite, alunite, and pyrite. The altered dacite contains approximately 49 per cent. quartz, 24 per cent. kaolinite, 16 per cent. alunite, and 7 per cent. pyrite, and the replacement has been attended by losses of much CaO, MgO, and Na_2O and of some SiO_2, Al_2O_3, and K_2O. Water, sulphuric acid, and sulphur have been added. There is no change in titanium and phosphorus. Under the assumption that no change in volume has taken place the altered rock has lost 5.3 per cent. in weight and now has a porosity of 10 per cent.

The ore-shoots form irregular bodies in the ledges and have no visible limits other than those indicated by the assay, but the ore is usually marked by more distinct brecciation and seaming than the surrounding ledge rock. During the earlier years little of the ore contained less than $30 per ton, but in 1916 ores yielding only $8 per ton were treated. Some shoots are roughly

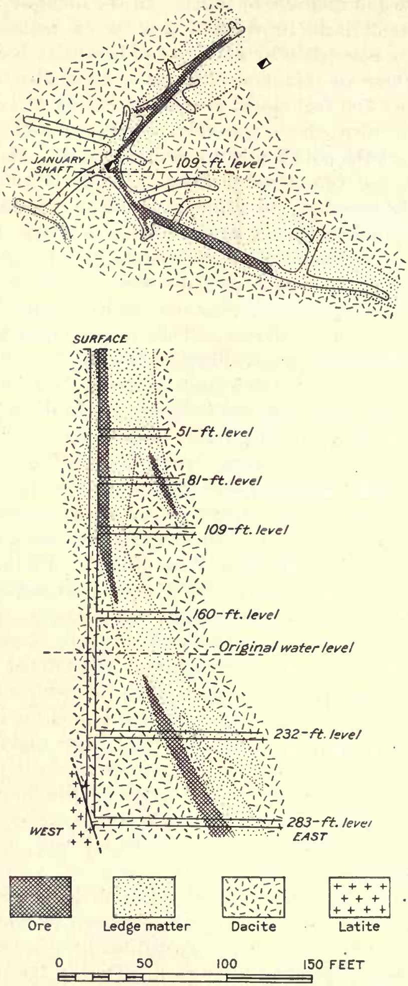

FIG. 181.—Plan of the 109-foot level and vertical section of the January mine, Goldfield, Nevada. *After F. L. Ransome, U. S. Geol. Survey.*

equidimensional masses; others are tabular, though often twisted and warped. In the lower levels (a depth of 1,750 feet has now been attained in the principal mine) the ore-bodies are said to be more veinlike and regular,[1] and the ore contains more copper. Much low-grade ore containing gold, copper and silver is said to have been found on the contact of latite and shale. The rich shoot in the Mohawk ground at a depth of 200 feet formed an irregular series of chambers about 100 feet high with a stope length of 75 feet; below the 245-foot level this shoot merges with other shoots to form an irregular chain of workings. A shoot in the Combination mine (Fig. 182) on the 130-foot level is 500 feet long, up to 40 feet wide, and 100 feet high. Ransome states that there is a gradual though irregular decrease in the tenor of the ore in depth and that at a depth of about 1,000 feet the workings are likely to pass into underlying rocks —andesite, or pre-Tertiary basement rocks—and this would probably be attended by changes in the mineralization. This prediction has been verified. While the shoots are not large compared to those of many other mines, the ore has been extremely rich. Much of it has averaged $419 per ton, containing 20

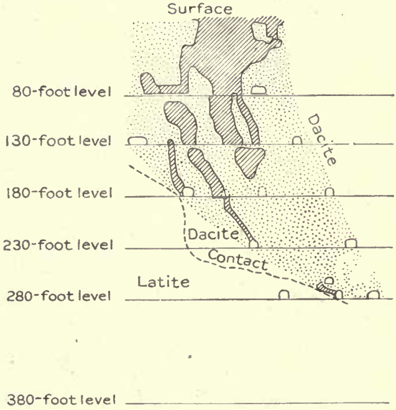

FIG. 182.—Vertical section of the Combination ledge, Goldfield, Nevada. *After F. L. Ransome, U. S. Geol. Survey.*

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[1] A. Locke, *Eng. and Min. Jour.*, Oct. 26, Nov. 2, 1912.

ounces of gold and 3 ounces of silver to the ton. One shipment of 14½ tons is said to have brought $45,783.[1]

The unoxidized ore contains fine-grained pyrite and marcasite, bismuthinite, goldfieldite ($5CuS.(Sb, Bi, As)_2(Se Te)_3$, an arsenical famatinite (Cu_3SbS_4), and native gold, also tellurides,[2] all in a dark-gray flinty quartz gangue (Fig. 183). Concentric shells of ore minerals with much finely divided yellowish-brown native gold about greatly altered fragments of rock are rather characteris-

FIG. 183.—Photomicrograph of silicified dacite, Goldfield, Nevada, showing quartz of varying grain. *p*, Pyrite; *t*, famatinite; *q*, quartz; *h*, vug. Magnified 40 diameters. *After F. L. Ransome, U. S. Geol. Survey.*

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[1] F. L. Ransome, *op. cit.*, p. 171.

[2] F. L. Ransome, *op. cit.*, pp. 115–116.

W. J. Sharwood, Gold tellurides, *Min. and Sci. Press*, vol. 94, 1907, p. 731.

tic of the rich ore. Other sulphides like galena and zinc blende, which elsewhere are common, are rare at Goldfield. The following analysis or rich ore shows well its unusual character.

ANALYSIS OF ORE FROM THE MOHAWK MINE, GOLDFIELD, NEVADA

	Per cent.		Per cent.
SiO_2	66.30	Bi	0.35
Al_2O_3	9.09	Te	2.42
CaO	none	Sb	trace
MgO	0.24	Au	2.00[1]
H_2O	7.00	Ag	0.25
Cu	2.08	S	6.30
Fe	3.83		
Zn	trace		99.86

[1] Equal to 541 fine ounces per ton.

While the deposition of pyrite, alunite, and kaolinite proceeded during the whole epoch of ledge formation, the richest ores were deposited somewhat later in brecciated and shattered parts. The water level stands 100 to 150 feet below the surface; above this the ores are oxidized and contain some gypsum but do not differ greatly in tenor from the fresh ores below water level.

These remarkable deposits are believed to have been formed by ascending alkaline waters, containing hydrogen sulphide, which derived their load of rare metals from deep-seated magmas. Through oxidation near the surface sulphuric acid was generated from the hydrogen sluphide and this acid attacked the rocks, causing the alunitic alteration.[1] The sulphuric acid, descending and mingled with surface waters, acted as a precipitant for the gold solution, and this combination of ascending alkaline and descending acid waters has, according to Ransome, probably resulted in the development of this unusual type.

The surface at the time of ore deposition was probably only a few hundred feet above the present surface.

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[1] Alunite is said to continue as a gangue mineral down to the low grade ore-bodies on the shale contact. These also contain much pyrite and famatinite but little free gold and tellurides.

CHAPTER XXV

METALLIFEROUS DEPOSITS FORMED AT INTERMEDIATE DEPTHS BY ASCENDING THERMAL WATERS AND IN GENETIC CONNECTION WITH INTRUSIVE ROCKS

GENERAL FEATURES

It is exceedingly common to find metalliferous deposits in or near bodies of intrusive rocks. These deep-seated rocks have been exposed by long-continued erosion and in places it is possible to arrive at a good estimate of the thickness of rocks removed, especially where the total thickness of the sedimentary series in which the intrusion occurred is known. That the deposits are not of recent development, but were formed a short time after the intrusion, can in most cases be proved conclusively, and from this it follows that they were developed under a great thickness of covering rocks. Their depth below the actual surface at the time of mineralization may, roughly speaking, be considered as ranging from 4,000 to 12,000 feet. In most of these deposits the absence of high-temperature minerals, such as magnetite, garnet, pyroxene, or tourmaline, shows that a high degree of heat did not prevail at the time of genesis. The depth below the surface indicates that the normal rock temperatures would be from 50° to 125° C., but in all cases the vicinity of recently intruded rocks had forced the temperature curves nearer to the surface; the heated waters which deposited the ores either emanated from the intrusive magma or at least derived their high temperatures from it. It is manifestly impossible to give accurate figures, but reasoning from what is known of the stability of minerals characteristic of this class of deposits we may say with some confidence that the actual temperatures may have ranged from 175° to 300° C. When the high-temperature curves were near the surface, these deposits may have originated at a depth of only a few thousand feet; when the intrusions were deeper seated the depth at which the deposits were formed may have exceeded 12,000 feet.

The pressure was necessarily strong; as calculated on a hydrostatic basis it ranged from 140 to 400 atmospheres. Communication with the surface was probably established in many places; where it was lacking the water and gases, propelled from the magma, may have been under still higher pressure.

When the temperature exceeded the upper limit stated above, silicate minerals characteristic of greater heat undoubtedly developed and the resulting deposit is of a different type. When the temperature fell below the lower limit stated, the general type of mineralization must have approached that of the deposits found near the surface.

The structure of the deposits is what might be expected from the opening of fissures under pressure at considerable depth. The fissures are fairly regular in strike and dip, having neither the extreme brecciated structure common to deposits formed close to the surface nor the lenticular form and irregular openings of the deep-seated deposits. Smooth walls and slickensides are abundant. As the fissures were opened in the zone of fracture, open spaces are present in many deposits, though the walls usually come together within short intervals. In calcareous rocks, more rarely in igneous rocks or quartzite, replacement deposits were often developed; they are more common here than in the deposits formed close to the surface, though the solutions, on the whole, spread much less widely through the igneous rocks in this group than in the shallow deposits.

The metals contained are principally gold and silver, often with large amounts of copper, lead, and zinc. In the deep-seated deposits molybdenum, bismuth, tungsten, and arsenic are not uncommon associates; we find the same metals here, though they are much less prominent; in addition there is also much antimony, and in places tellurium. The ore minerals are sulphides, arsenides, sulphantimonides, and sulpharsenides. Pyrite, chalcopyrite, arsenopyrite, galena, zinc blende, tetrahedrite, tennantite, and native gold are the most common and on the whole there is not much variety and complexity. Scarcely ever do we find the oxides such as magnetite, specularite, and ilmenite. The metallic minerals develop both in the filling and in the altered country rock, but in the fissure veins proper it is common to find the valuable ores mainly in the filled spaces. The predominating gangue mineral is quartz, but carbonates are also common, such as calcite, dolomite, and ankerite, more rarely siderite; fluorite and barite are occasionally of importance. Chalcedony and opal are rarely found.

Among the minerals of this type of metallization are found no biotite; no pyroxenes or amphiboles; no garnet, tourmaline, or topaz; no zeolites or kaolin.

Very frequently these veins follow lamprophyric dikes, which are usually the last manifestation of igneous activity.

This class yields a large proportion of the gold production of the world, as well as much of its silver, copper, and zinc. Its deposits are by far the most abundant in the Cordilleran region, as well as in other parts of the world where intrusive activity has been followed by deep erosion.

The gold-quartz veins of California and of Victoria (Australia) and many of those of the Cordilleran region, the zinc-lead-silver replacement deposits at Leadville, Park City, and Aspen, the Coeur d'Alene lead veins, and many other types belong to this class.

The intrusive bodies may be laccoliths, stocks or batholiths; the latter term being reserved for intrusive cross-cutting masses of large size like those of the Helena-Butte region in Montana, the great Idaho granite mass, or the enormous intrusive masses following the Pacific Coast from lower California to Alaska. It has long been known that the interior parts of great batholiths are relatively barren and that mineral deposits rarely occur in laccoliths. This appears to be dependent upon the principles of magmatic differentiation for the volatile constituents of the magma from which the ore deposits have been formed have a tendency to rise to the roof of the intrusive or to accumulate in the cupolas of its upper parts.[1]

Attention has recently been directed to these problems by Butler,[2] who points out that in Utah those stocks which are truncated by erosion near their apices or points nearest to the surface contain ore deposits of great importance while those which have been truncated by erosion to deeper levels show relatively few deposits. When the mobile constituents reached a point when the magma was sufficiently consolidated to fracture they were guided by the fissures and on reaching favorable physical and chemical environments began to deposit the metals in solution.

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[1] R. A. Daly, Igneous rocks and their origin, New York, 1914, p. 244 and 253.

[2] B. S. Butler, Relation of ore deposits to different types of intrusive bodies, *Econ. Geol.*, vol. 10, 1915, pp. 101–122.

Billingsley and Grimes[1] have lately applied the same theory to the Boulder Batholith, Montana, and appear to have shown that the maximum deposition took place near the roof of that batholith.

Many veins of this class show interesting changes in depth and laterally from the focus of the ascending waters. In general the ore is likely to grow more siliceous and pyritic in depth though many cases are known where there is practically no change in the composition for a vertical distance of several thousand feet. Gold, copper and zinc appears to have been deposited by relatively hot waters while lead and rich silver deposits as a rule appear farther away from the intrusive suggesting deposition at lower temperatures (p. 190).

METASOMATIC PROCESSES

General Character.—The alteration of the country rock is usually very intense next to the ore, but seldom yields coarsely crystalline products as in some high temperature deposits (p. 659). In feldspathic and ferromagnesian rocks the principal product is sericite, the fine-grained foliated form of muscovitel in many deposits carbonates, such as calcite, dolomite, and ankerite, develop in large amounts. The dark minerals are first altered, their iron being usually recombined as pyrite. The feldspars are also rather easily altered; quartz grains are attacked, though less easily than the feldspars, and are partly, at least, converted to a similar aggregate of sericite and carbonates.

The granodiorite adjoining a gold-quartz vein may be altered, for instance, to a rock composed of 16 per cent. quartz, 42 per cent. sericite, 33 per cent. calcium (magnesium, iron) carbonate, and 9 per cent. pyrite. While the orthoclase and the soda-lime feldspars are conspicuously absent as vein-forming minerals, albite is not uncommon, especially in some gold-quartz veins. This mode of alteration is frequently observed in amphibolitic rocks, which contain much sodium and presumably much albite developed during previous dynamo-metamorphism. In many cases the new albite and dolomitic or ankeritic carbonates form together. Pyrite is a common metasomatic mineral and is often

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[1] P. Billingsley and J. A. Grimes. The ore deposits of the Boulder Batholith, *Trans.*, Am. Inst. Min. Eng., vol. 58, 1918, pp. 284–368.

associated with the ferromagnesian minerals, but may also form in quartz and feldspars. Other metallic minerals are not common; the apatite and zircon of the igneous rocks resist alteration; while titanite and ilmenite yield rutile. In many vein types of the interior Cordilleran province the metasomatic carbonates are scarce or absent, as in the copper veins of Butte, Montana, and Clifton, Arizona. Serpentine is sometimes altered to magnesite and dolomite.

Among sedimentary rocks quartzite and sandstone are little affected, except in veins of the Coeur d'Alene type, where the quartz grains are replaced by siderite. Clay slates always contain metasomatically developed pyrite in cubes; whether they are otherwise altered or not depends upon their composition: if they contain feldspathic sediment, sericitic and carbonatic alteration will ensue; if only kaolin, sericite, and quartz are present there will be little noticeable alteration, except in some instances where almost complete silicification takes place.

Limestone and other calcareous rocks are almost always subject to silicification by the replacement of the carbonates with fine-grained quartz aggregates; the resulting rocks are usually called "jasperoids" and look more or less like chert (Figs. 57 and 58). Ore minerals develop abundantly by metasomatic action in such rocks.

Along with or preceding this silicification, dolomitization often takes place; the solutions apparently abstract a part of the calcite in the limestone and replace it by the less easily soluble magnesium carbonate. Limestone is sometimes converted to magnesite, siderite, manganosiderite or fluorite.

The alteration is accompanied by strong leaching of sodium and by concentration of potassium. Where there is little carbonatization much calcium and magnesium are also leached. Aluminum in most cases remains about constant.

Alteration of Wall Rocks Adjoining Gold-Quartz Veins.—In veins characterized by quartz filling with free gold and simple sulphides or arsenopyrite, the country rock next to the walls is usually rich in carbonates, sericite, and pyrite, but rarely contains much gold. Extensive alteration zones are not common, and sometimes fresh rock adjoins the vein. The relative quantity of alteration products may differ considerably, even in the same mine.

Clay slate, with more or less carbonaceous matter, is thought by some to have a precipitating and enriching influence on the vein, but to what extent this is true is doubtful. While this influence can apparently be recognized in some districts like Gympie, in Queensland, it is not clearly shown along the Mother Lode of California. The black clay slates near the veins are often rich in crystallized cubes of pyrite.

Metasomatic rocks containing albite result from the alteration of amphibolites at Angels Camp, Calaveras County, California, where they constitute low-grade ores. The Utica, Lightner, and Melones mines are the best known of those working on these deposits. The metasomatic processes were similar to those that affected the amphibolites of Kalgoorlie. Ransome[1] has described the Angels Camp deposits.

Specimens from Angels Camp and Melones show that some of the altered rocks consist of sericite with embedded grains of calcium-magnesium-iron carbonates and pyrite. In other places the carbonates prevail over the sericite, while in a third and very common type much of the sodium, abundant in the amphibolite, has been retained as albite in the altered rock. Large grains of carbonates are separated by a granular mass of quartz, albite, pyrite, and sericite.

In the ordinary course of alteration the ferromagnesian minerals are first converted into larger foils of sericite. A chlorite rich in iron is also formed, which during a later stage is converted into sericite. The feldspars are then attacked along cracks and cleavage planes, and a finely felted aggregate of sericite and calcite invades the grains until the replacement is complete. A texture often observed consists of interlacing sericite foils, the triangular or polygonal interstices of which are filled with calcite (Fig. 184). Orthoclase is almost always more resistant than the soda-lime feldspars.

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[1] F. L. Ransome, *Folio* 63, U. S. Geol. Survey. See also W. Lindgren, *Econ. Geol.*, vol. 1, 1906, p. 543. Of these veins, Ransome says:

"They are in the main complex stringer leads. But the country rock is usually much more altered and may be heavily impregnated with pyrite near the vein. It is often changed to a soft grayish rock consisting chiefly of carbonates of lime and magnesia, with sericite and sometimes a little chlorite. Such altered and pyritized country rock is too poor in gold to pay for working alone, but is often run through the mills for the sake of the rich stringers which intersect it. These veins are usually richer in carbonates than those in the black slate areas and in certain parts of the district are rich in tellurides."

The quartz is also attacked, but with more difficulty, and is in no case completely replaced. Magnetite appears to be converted to siderite and titanite to rutile. A part of the ferromagnesian minerals are transformed into pyrite Sharp cubes of pyrite develop, however, not only in the sericitic aggregate, but also in the fresh feldspars or even in the quartz. Arsenopyrite is almost the only other sulphide which is enabled to form in the altered rock, and it develops in sharp rhombic crystals.

FIG. 184.—Altered granodiorite, Bellefountain mine, Nevada City, California. *m*, Fine aggregate of sericite with some calcite and quartz, replacing orthoclase and andesine; *b*, original biotite altered to sericite; *q*, original quartz; black, pyrite with included sericite. Magnified 15 diameters.

From many analyses the eight given on page 553 are selected.[1]

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[1] W. Lindgren, The gold-silver veins of Ophir, Cal., *Fourteenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1894, pp. 243–284.

W. Lindgren, Metasomatic processes in fissure veins, *Trans.* Am. Inst. Min. Eng., vol. 30, 1900, p. 666.

TABLE I.—ANALYSES OF METASOMATIC ROCKS FROM CALIFORNIA GOLD QUARTZ VEINS

	A	A_1	B	B_1	C	C_1	D	D_1
SiO_2	65.54	46.13	45.56	37.01	66.65	56.25	51.01	45.74
TiO_2	0.39	0.67	1.11	0.85	0.38	0.25	0.98	0.36
Al_2O_3	16.52	15.82	14.15	12.99	16.15	17.65	11.89	5.29
Fe_2O_3	1.40	0.89	1.20	0.43	1.52	0.76	1.57	0.13
FeO	2.49	2.27	9.83	3.57	2.36	2.64	6.08	2.06
FeS_2	....	1.61	7.86	7.99	0.02	2.87	[1]1.73	0.49
Cu_2S	....	.....	0.10	trace	......	......	trace	......
MnO	0.06	0.09	0.25	0.24	0.10	none	trace	0.26
NiO, ZnO	....	trace	trace	trace	......	......	......	......
CaO	4.88	10.68	2.30	9.78	4.53	4.46	10.36	23.85
SrO	not det.	trace	trace	trace	trace	......	none	none
BaO	not det.	trace	trace	trace	0.07	0.03	none	trace
MgO	2.52	2.13	6.76	5.49	1.74	1.69	8.87	0.94
K_2O	1.95	5.30	1.18	4.02	2.65	6.01	0.15	1.29
Na_2O	4.09	0.17	1.57	0.13	3.40	0.30	4.17	0.11
Li_2O	....	trace	trace	trace	trace	......	......	trace
H_2O-	....	0.12	0.23	0.13	0.18	0.30	0.24	0.22
H_2O+	0.59	2.42	4.84	1.92	0.72	2.36	2.09	1.07
P_2O_5	0.18	0.10	0.14	0.06	0.10	0.21	0.17	0.07
SO_3	....	.....	0.03	0.04	......	......	......	......
CO_2	....	11.24	3.04	15.04	......	4.82	......	18.91
Total	100.61	99.64	100.15	99.69	100.57	100.60	99.35	100.79

[1] Probably present as Fe_7S_8.

A. Fresh granodiorite, Lincoln, Placer County, California. Though not adjoining the vein, it indicates closely the actual composition of the fresh wall rock. W. F. Hillebrand, analyst.

A_1. Altered granodiorite, Plantz vein, Ophir, Placer County. W. F. Hillebrand, analyst.

B. Amphibolite schist, Conrad vein, Ophir, Placer County. Fairly fresh, but contains pyrite and calcite. W. F. Hillebrand, analyst.

B_1. Completely altered amphibolite schist, Mina Rica vein, Ophir, Placer County. W. F. Hillebrand, analyst.

C. Fresh granodiorite, Nevada City, Nevada County. W. F. Hillebrand, analyst.

C_1. Altered granodiorite, Bellefountain mine, Nevada City. George Steiger, analyst.

D. Fresh diabase, Grass Valley. H. N. Stokes, analyst.

D_1. Altered diabase, North Star mine, Grass Valley. W. F. Hillebrand, analyst.

From the chemical and microscopical data the following compositions may be calculated:

TABLE II.—CALCULATED MINERALOGICAL COMPOSITION OF THE ALTERED ROCKS OF TABLE I

	A_1	B_1	C_1	D_1
Quartz	16.00	24.00	25.00	35.00
Sericite (with a little chlorite)	41.76	46.97	61.46	21.20
$CaCO_3$	17.53	18.87	7.23	42.15
$MgCO_3$	9.67	2.93	2.70	0.71
$FeCO_3$	5.76	3.67	0.58	....
$MnCO_3$	0.42	0.14	....	....
Rutile	0.85	0.67	0.25	0.36
Pyrite	7.99	1.61	2.87	0.50
Apatite	0.13	0.22	0.46	0.15
Total	100.11	99.08	100.55	100.07

As it seems probable that the alumina has remained fairly constant in the first six analyses in Table I, they may be directly compared for an approximate review of the chemical changes.

Analysis D_1 differs from the rest in showing an exceptionally high percentage of introduced lime and carbon dioxide and a corresponding loss of magnesia. Moreover, the alumina is so low that it must be supposed to have been removed.

The characteristic features of the process seem to consist in the decrease of silica, magnesia, and soda and increase of lime, potash, and carbon dioxide—the calcitic altered rock strongly contrasting with the quartz-filled veins. There is some evidence of partial leaching of titanium and phosphorus. Sufficient data are not available for the accurate determination of change of volume during the process, and of the actual losses and gains. It seems probable that, in most cases, the added material has more than balanced the losses. Unquestionably there has been a strong addition of calcium and potassium, and the vein-filling process probably began with deposition by solutions extremely rich in these constituents, as well as in carbon dioxide. The quartz filling sometimes shows imprints, along its walls, of calcite crystals, from which it may be concluded that during the process of filling the nature of the solutions changed to the later phase, in which almost nothing but quartz was deposited.

Interior Types.—In the Blue Mountains of Oregon the wall rocks of the veins are altered to products in all respects similar to those of California. In the veins of Idaho, Utah and Colorado genetically connected with Cretaceous or early Tertiary intrusions of quartz monzonite or similar porphyries, the carbonatization is far less marked and both calcium and mangnesium are leached. The accompanying analyses, Table III, illustrate the chemical changes in two prominent types.

TABLE III.—ANALYSES OF FRESH AND ALTERED ROCKS FROM IDAHO GOLD-QUARTZ VEINS

[Analyst, W. F. Hillebrand]

	E	E_1	F	F_1
SiO_2	65.23	66.66	57.78	58.01
TiO_2	0.66	0.49	1.01	1.08
Al_2O_3	16.94	14.26	16.28	15.72
Fe_2O_3	1.60	0.67	1.02	0.64
FeO	1.91	1.33	4.92	3.87
CoO, NiO	....	....	0.02	0.12
MnO	trace	trace	0.15	0.17
CaO	3.85	3.37	6.65	2.15
SrO	....	....	0.07	none
BaO	0.19	none	0.12	trace?
MgO	1.31	0.95	4.60	2.07
K_2O	3.02	4.19	2.22	4.79
Na_2O	3.57	none	3.25	0.10
H_2O-	0.18	0.36	0.34	0.31
H_2O+	0.88	2.16	0.92	2.71
P_2O_5	0.19	0.17	0.30	0.31
CO_2	0.25	3.67	0.15	2.86
S	none	0.95	0.02	1.25
Fe	....	0.84	....	1.52
Pb	....	....	....	0.86
Cu	....	....	....	0.05
As	....	....	....	1.65
Total	99.78	100.07	99.82	100.24

E. Fresh granitic rock immediately adjoining the Silver Wreath quartz vein, Willow Creek, Idaho.

E_1. Altered rock adjoining the same vein.

F. Fresh quartz-pyroxene diorite adjoining the Croesus vein, Hailey, Idaho.

F_1. Altered rock adjoining the same vein.

E and E_1 represent the fresh and altered rock from the Willow Creek district, Boise County, where the narrow quartz veins carry scarcely any free gold, but much auriferous galena, pyrite, arsenopyrite, and zinc blende. F and F_1 represent the fresh and altered rock from the Croesus mine, Wood River district, Blaine County, where the narrow streaks of filling consist of quartz siderite, pyrrhotite, and chalcopyrite, with a little galena, arsenopyrite, and zinc blende. Here again only a fraction of the gold is in the free state. The ore contains very little silver.[1]

The specific gravity of E is 2.714. From the mineralogical composition given in the report quoted the specific gravity is calculated to 2.720, which is a close agreement, the difference possibly indicating a very slight porosity.[2]

TABLE IV.—MINERALOGICAL COMPOSITION OF E_1 AND F_1, IN TABLE III

	E_1	F_1
Quartz	42.00	36.18
Sericite	46.84	38.18
Chlorite	....	11.76
$CaCO_3$	4.80	3.11
$MgCO_3$	1.96	1.26
$FeCO_3$	1.45	2.19
Rutile	0.49	1.08
Apatite	....	0.72
Pyrite	1.78	0.58
Pyrrhotite	....	0.15
Zinc blende	....	trace
Galena	....	0.99
Chalcopyrite	....	0.15
Arsenopyrite	....	3.58
Total	99.32	99.93

The measured specific gravity of E_1 is 2.774, indicating that the rock alters to denser minerals. The calculation of the same

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[1] For full calculations and description of E and E_1 see W. Lindgren, *Eighteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1898, p. 640; for F and F^1 see W. Lindgren, *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, pp. 211–232.

[2] In this calculation the following figures for specific gravity are used: quartz, 2.65; sericite, 2.83; biotite, 3.00; oligoclase, 2.65; orthoclase, 2.56.

specific gravity Table from IV gives 2.796, which shows a decided porosity of the altered rock. Under these circumstances, no evidence of pressure being noted, it may be assumed that not much change in volume has taken place. By multiplying the percentages of E and E_1 by 2.714 and 2.774, respectively, and comparing the results, the absolute gains and losses per cubic meter may be obtained (see Table V).

In the same manner the measured specific gravities of F and F_1 are compared with the calculated specific gravities.[1]

By multiplying the percentages of F and F_1 by the measured specific gravities, and comparing these data, the absolute gains and losses are again obtained.

The calculation shows that during the alteration of E to E_1 291 kilograms were added and 229 lost per cubic meter, the net gain being 62 kilograms. During the alteration of F to F_1 416 kilograms were added and 333 lost per cubic meter, the net gain being 83 kilograms.

Similar changes resulted in the two rocks: a moderate addition of silica and a strong gain of potash; nearly complete loss of soda, baryta, and strontia; partial loss of alumina, magnesia, and lime, F, however, losing much more lime than E. In E_1 the amounts lost of Fe_2O_3 and FeO are almost completely converted into Fe (in FeS_2). In F these losses are less and not sufficient to account for the gain of Fe; consequently iron must have been added. Phosphoric acid is constant, consistently with the fresh state of the apatite.

These figures give some idea of the intensity of the transfer of material, though the balance of material added is comparatively small.

Similar alteration in the veins of Bingham, Utah, has been described by Boutwell.[2] The Last Chance lode is from 1 to 14 feet wide and contains galena, zinc blende, pyrite, and some calcite. The alteration and bleaching extend about 1 or 2 feet into the country rock, which is a monzonite, consisting of orthoclase, plagioclase, augite, biotite, and hornblende. The alteration begins by chloritization and dissemination of pyrite, but the end product consists largely of sericite and pyrite. The analysis indicates an unusual and almost complete removal of magnesia and extensive leaching of sodium and calcium. There has

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[1] *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, pp. 221–232.

[2] J. M. Boutwell, Bingham mining district, *Prof. Paper* 38, U. S. Geol. Survey, 1905, p. 178.

evidently been an addition of potassium, as there is considerably more than is called for in the ordinary composition of sericite. As usual TiO_2 remains constant, and the altered rock contains practically no carbonates.

ANALYSES SHOWING ALTERATION OF MONZONITE AT BINGHAM, UTAH

[Analyst, E. T. Allen]

	I	II
SiO_2	58.64	56.78
TiO_2	0.83	0.81
Al_2O_3	15.35	16.90
Fe_2O_3	3.25	6.87
FeO	2.54	2.34
CaO	5.37	1.18
BaO	0.18	0.14
MgO	3.84	0.03
K_2O	4.23	7.02
Na_2O	3.60	0.37
H_2O-	0.86	1.32
H_2O+	1.50	2.23
CO_2	none	0.26
P_2O_5	0.02	0.04
S	0.05	5.93
	100.26	102.22
O equivalent to S	0.02	2.22
Total	100.24	100.00

Traces MnO, Cr_2O_3.

I. British Tunnel, Last Chance mine.

II. British Tunnel, Last Chance mine, at wall of lode.

One of the great mineral belts of Colorado extends in a northeasterly direction from Leadville to Boulder by way of Park, Clear Creek, and Gilpin counties (p. 617). It is characterized as a whole by an abundance of heavy sulphide ores, principally pyrite, zinc blende, and galena, with subordinate chalcopyrite and a notable content of gold and silver. The gangue is subordinate and consists of a little quartz and more or less of a sideritic carbonate. The ores appear in replacement deposits and veins. At Leadville, where the ores replace limestone at the contacts with intrusive porphyry, the alteration of the carbonate rock is remarkably slight. There may be locally a little pyrite in the limestone near the ore; at other places the limestone is recrystallized and contains much manganosiderite or is silicified. The contact between ore and limestone is usually very sharp—indeed, in spite of the completeness of the replacement, practically unaltered limestone may lie next to the ore.

At Breckenridge, Georgetown, and Central City the deposits are fissure veins, generally filled with massive sulphides, and, in feldspathic rocks, they are adjoined by an altered zone from a few inches to 20 feet or more in width. The alteration products are quartz, sericite, and a sideritic carbonate, with more or less pyrite. In a few of the Georgetown veins Spurr[1] detected adularia, but, on the whole, this mineral is absent.

The siderite is thought to have been derived from biotite and magnetite, and the gangue minerals are believed to have been derived from the adjoining rocks. Kaolin is considered to have resulted from alteration by descending waters during the processes of weathering and sulphide enrichment.

The course of alteration near the veins has been studied in detail by Ransome[2] at the Wellington lode, 350 feet below the surface and below the zone of oxidation. The vein is here 5 feet wide with good walls and contains zinc blende and galena in a little gangue of siderite and barite with more or less included country rock. The alteration spreads 20 feet from the vein. The fresh rock is a dark-gray monzonite porphyry, the groundmass of which consists of labradorite, orthoclase, biotite, and diopside. The altered rock is light gray, with disseminated particles of sulphides. The rock, while retaining a faint trace of its structure, is changed to a mass of carbonate, sericite, and quartz.

By multiplying the figures of the percentage composition by the specific gravities of the rock mass the constituents per 100 cubic centimeters of fresh and altered rock are obtained. These figures compared give the gains and losses for each constituent during the alteration of 100 cubic centimeters of fresh rock, and from these again may be calculated the gains and losses in percentage of the original mass of 276.3 grams of fresh rock. These gains or losses in percentages may be applied directly by addition

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[1] J. E. Spurr, Economic geology of the Georgetown quadrangle, *Prof. Paper* 63, U. S. Geol. Survey, 1908.

[2] F. L. Ransome, The Breckenridge district, *Prof. Paper* 75, U. S. Geol. Survey, 1911, pp. 95–101.

or subtraction to the figures of the chemical analysis of fresh rock, and this, as shown in column III, will express the nature of the change more clearly. There has been a notable loss of silica, calcium, potassium, and sodium. The additions comprise carbon dioxide, sulphides, ferrous iron, and magnesium, which would hardly bear out Spurr's assertion that the siderite in the vein is derived from the adjoining country rock. As usual apatite remains unaltered, and the ilmenite is converted to rutile. Some paragonite has probably developed besides the sericite, if indeed the rock does not contain albite.

TABLE SHOWING ALTERATION OF DIORITE PORPHYRY AT BRECKENRIDGE, COLORADO

	I	II	III
SiO_2	57.35	46.62	49.48
TiO_2	1.07	1.01	1.07
Al_2O_3	16.29	12.66	13.44
Fe_2O_3	3.15	trace	0.02
FeO	4.36	11.15	11.78
MnO	0.12	0.92	0.97
CaO	5.66	1.55	1.66
BaO	0.10	none	....
SrO	0.05	none	....
MgO	2.41	4.02	4.25
K_2O	3.39	1.68	1.79
Na_2O	4.50	1.35	1.45
H_2O-	0.15	0.31	0.33
H_2O+	0.70	3.41	3.60
CO_2	0.46	11.48	12.11
P_2O_5	0.70	0.50	0.53
FeS_2	0.09	1.99	2.10
ZnS	none	0.97	1.02
PbS	....	0.52	0.55
Total	100.55	100.14	106.15
Specific gravity:			
Mass	2.763	2.930	
Powder	2.799	2.940	

I. Diorite porphyry, 25 feet from vein, Wellington mine.

II. Altered porphyry, close to vein, Wellington mine.

III. Composition of same volume of altered rock in percentage of original rock mass.

An approximate calculation shows the altered rock to be composed as follows:

APPROXIMATE MINERALOGICAL COMPOSITION OF ALTERED DIORITE PORPHYRY AT BRECKENRIDGE, COLORADO

Sericite	30.5	Apatite	1.3
Quartz	31.6	Pyrite	2.0
Kaolinite	2.8	Sphalerite	1.0
Carbonate	29.3	Galena	0.5
Rutile	1.0	Total	100.0

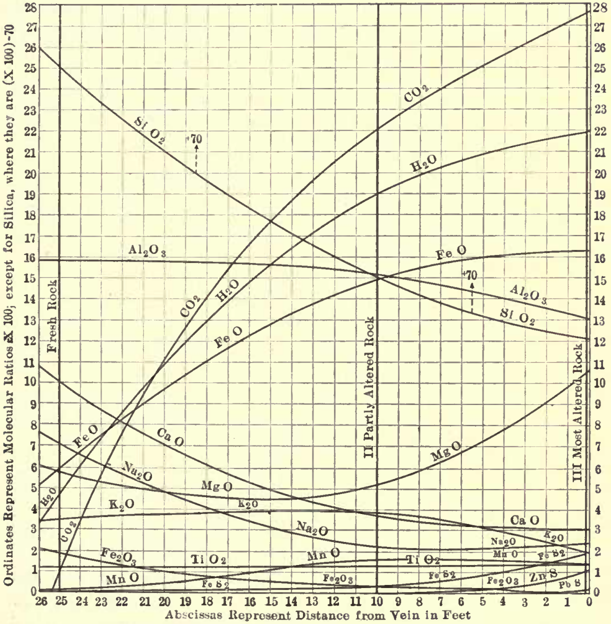

FIG. 185.—Diagram showing alteration of diorite porphyry by vein-forming solutions at Breckenridge, Colorado. *After F. L. Ransome, U. S. Geol. Survey.*

The carbonate consists of 63.9 per cent. $FeCO_3$, 29.6 per cent. $MgCO_3$, 5.2 per cent. $MnCO_3$, and 1.3 per cent. $CaCO_3$, all in isomorphous mixture.

The majority of the large copper deposits of the West are genetically connected with intrusives and have been formed at intermediate depths, though some of them like Ely and Clifton and Cactus show unmistakable affiliations with the high temperature deposits. The thermal alteration of the feldspathic rocks results universally in sericite, quartz and pyrite with practically no carbonates. Omitting minor constituents the composition of the altered rocks would average, in per cent., about 65.0SiO_2, 16.0Al_2O_3, 2.0FeO, 0.5Fe_2O_3, 1.0MgO, 0.25CaO, 0.5 Na_2O, 5.0K_2O and 5.0FeS_2.[1]

From all this it follows that the mineralization in mineral deposits of this class has been effected by solutions of relatively uniform character capable of substituting K_2O for Na_2O. In some deposits, particularly those carrying gold, silver and lead the alkaline earths have been fixed as carbonates, while in copper deposits there is usually no carbonates in the metasomatic products.

The total result indicates action by hot ascending solutions containing more potash than soda and having a variable amount of alkaline carbonates and free carbon dioxide.

PARAGENESIS

It has long been observed that the minerals are formed, in the main, in an orderly succession, which sometimes is repeated. This was first emphasized by Breithaupt, who recorded the series for different mineral deposits in his book on the paragenesis of minerals. The study of the succession has an evident bearing on scientific and economic problems, and much work has been done lately on this subject in connection with the examination of ores in polished sections by metallographic methods. These have disclosed the wonderful extent to which metallic minerals are replaced by others during the process of metallization.

When both filling and replacement have been in action it is natural that the country rock would first be attacked by the waters. There was first a process of dialysis by means of which the solutions were filtered through a porous rock which many elements found it difficult to penetrate. Thus we find that the composition of the metasomatic rocks near the fissure differs

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[1] Clifton-Morenci, W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey; Butte, W. H. Weed, *idem,* 74; San Francisco district, Utah, B. S. Butler, *idem,* 80; Ely, A. C. Spencer, *idem,* 96; Ray and Miami, F. L. Ransome, *idem,* in press.

considerably from that of the filling. In gold quartz veins, for instance, there is as a rule little gold in the metasomatic rocks while the filling may be rich. The carbonates, pyrite, and sericite in the rock seem to have developed about contemporaneously. I am unable to accept Rogers'[1] suggestion that sericite is a mineral of late origin.

In general the arsenopyrite and pyrite appear early in the vein filling and are preceded and followed by quartz. The latter mineral nearly always appears in several generations. Calcite, dolomite and siderite are usually the latest gangue minerals. Chalcopyrite and bornite are always later than the pyrite and galena is one of the latest of the simple sulphides (Fig. 186). At times it crystallizes with the zinc blende. Argentite included in galena is evidently of simultaneous formation. Gold is frequently later than the earlier quartz generations, and replaces pyrite and arsenopyrite.

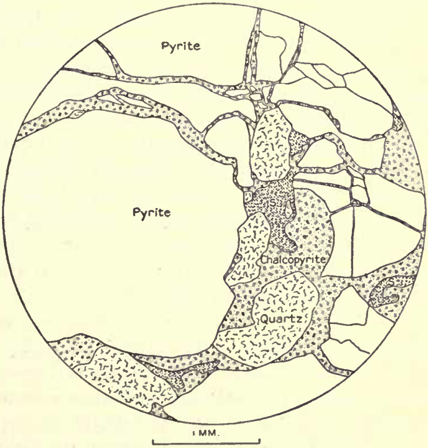

Fig. 186.—Drawing of polished surface of ore from Gilpin County, Colorado, showing earlier pyrite traversed by later veins of chalcopyrite, sphalerite (*s*), and quartz. *After E. S. Bastin.*

The sulphantimonides and sulpharsenides are almost invariably

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[1] A. F. Rogers, Sericite, a low temperature hydrothermal mineral, *Econ. Geol.*, vol. 11, 1916, pp. 118–150.

the last metallic minerals to form during the primary metallization. The occurrence of jamesonite, tetrahedrite, pearcite and ruby silver in vugs and replacing veinlets is a common observation. The periods of deposition of the various minerals frequently overlap and recur.

GOLD-QUARTZ VEINS OF THE CALIFORNIA AND VICTORIA TYPE

Principal Characteristics.—As quartz and gold may be deposited together within a considerable range of temperature, there are several types of gold-quartz deposits. The deposits formed at higher temperatures, distinguished by such gangue minerals as tourmaline, apatite, garnet, biotite, and amphibole, will be described in subsequent pages. Those formed near the surface at temperatures not much above 100° C. have been discussed in the preceding chapter. Between the two kinds stands the large group of important deposits whose geological relations point to development at considerable depth and whose mineral association points to moderate temperatures—perhaps 200° to 300° C.

The first type is represented by certain Appalachian and Brazilian gold-quartz veins; the second by veins like those of the Comstock, Bodie, and Tonopah, usually appearing in Tertiary lavas. Between the two stand the gold-quartz veins of California, eastern Australia, and many localities in the interior Cordilleran region of North America.

The general characteristic of these intermediate deposits is the association of a preponderant gangue of milky, coarsely crystalline quartz, sometimes drusy, though rarely showing comb structure, with free gold and auriferous simple sulphide minerals. Where the country rock is suitable for replacement carbonates and sericite appear with pyrite in the altered rocks.

The veins occur in deeply eroded regions and in or surrounding intrusives of quartz monzonitic or dioritic or gabbroitic kind. The absence of biotite, magnetite, epidote, garnet, and tourmaline is also notable. The only silicates present are albite and chlorite, and these only locally. The destruction of the outcrops by erosion usually results in rich placers, in which large nuggets of gold are often found.

The free gold always contains a little silver, the average fineness being 0.800; the sulphides are likely to carry more silver in proportion than the native gold. Some types of these veins carry a notable amount of silver, but scarcely ever such amounts as are common in the Tertiary veins deposited near the surface in Tertiary lavas.

Gold-Quartz Veins of the Sierra Nevada.[1]—The greatest development of the gold-quartz veins is found in California; they begin in the southern end of the State in San Diego County and continue with interruptions to the northern end, where, in Trinity and Siskiyou counties, there is a productive area of no small value. The gold belt also continues into southwestern Oregon, but farther north disappears under the Tertiary lavas and Cretaceous and Tertiary sediments.

Most typically the veins are developed in the Sierra Nevada, which, with its gentle western slope and abrupt eastern escarpments, separates the deserts of the Great Basin from the central valleys of California (Fig. 187).

The crest and main mass of this range form parts of an enormous batholith of massive granodiorite and allied rocks, intruded into Mesozoic and Paleozoic metamorphosed sediments. These sedimentary rocks are closely folded and compressed and occupy a belt on the western slope, which gradually widens and, in Plumas County, spreads over a width of 60 miles. The great batholith itself contains extremely few quartz veins; mineralization is confined to the belt of metamorphic rocks on the western lope and often begins abruptly at the contact; this is shown, for instance, by the river gravel, which becomes auriferous where

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[1] H. W. Fairbanks, Geology of the Mother Lode region, *Tenth Rept.*, California State Min. Bur., 1890, pp. 23–90.

W. Lindgren, Characteristic features of the California gold-quartz veins, *Bull.*, Geol. Soc. Am., vol. 6, 1896, pp. 221–240.

W. Lindgren, Gold-silver veins of Ophir, *Fourteenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1893, pp. 243, 284.

W. Lindgren, The gold-quartz veins of Nevada City and Grass Valley, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1896, pp. 1–262.

F. L. Ransome, The Mother Lode district, *Folio* 63, U. S. Geol. Survey.

W. H. Storms, The Mother Lode region of California, *Bull.* 18, California State Min. Bur., 1900.

W. H. Storms, Pocket mines, *Min. and Sci. Press,* June 6, 1908; *idem,* Possibilities of the Mother Lode in depth, Nov. 18, 1911.

W. H. Storms, The occurrence of gold at intersections, *Min. and Eng. World,* Nov. 25, 1911.

H. W. Turner and F. L. Ransome, Sonora, *Folio* 41, U. S. Geol. Survey, and other folios of the same region.

Charles G. Yale, Mine production of California, *Mineral Resources,* U. S. Geol. Survey, annual publication.

the streams enter the metamorphic areas. The highly productive part of the belt does not, usually, adjoin the granitic rocks, but appears lower down in the foothill region near smaller intrusive areas.

The metamorphic rocks are a complex body, for besides the prevailing Paleozoic slates with occasional lenses of limestone and in the lower foothills a narrow belt of late Jurassic Mariposa slate, they contain Paleozoic lava flows and a vast quantity of tuffs, diabases, and old andesites erupted by volcanoes of Jurassic age.

FIG. 187.—Map of Nevada and northern part of California, showing prominent mining districts.

Later than these rocks, and probably dating from earliest Cretaceous time, are numerous smaller intrusions of gabbro, diorite, and granodiorite, which are massive and, in a general way, contemporaneous with the main batholith of the range. The basic intrusions appear to be somewhat older than those containing more silica.

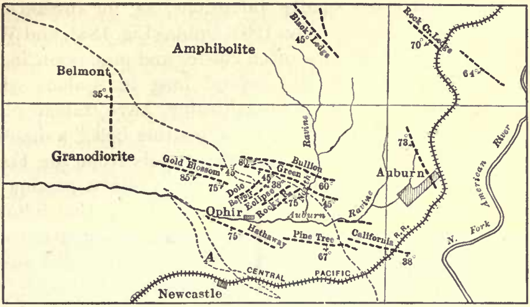

FIG. 188.—Map of principal vein systems near Ophir and Auburn, California. A, Small area of amphibolite. Scale 1 inch = 2.7 miles.

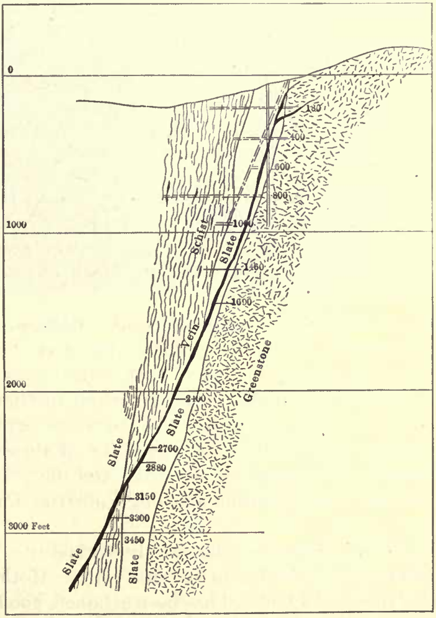

FIG. 189.—Vertical cross-section of the Mother Lode near the Argonaut shaft, showing reverse fault along vein fissure. "Schist" is amphibolite, white area, Mariposa clay slate (Jurassic).

In and around these smaller intrusions, as, for instance, at Grass Valley, Nevada City (Fig. 190), Ophir (Fig. 188), and West Point, the gold-bearing veins often cluster and may occur in any kind of rocks; there are also several long lines along which fracturing and subsequent mineralization have taken place. One of these follows the so-called "serpentine belt," a dike-like intrusive mass 70 miles long; another extends from the Forest Hill divide, in Placer County, up into Sierra County, passing the town of Washington. The most important line is that followed by the Mother Lode, in the foothills of Mariposa, Tuolumne, Calaveras, Amador, and Eldorado counties, for a distance of 130 miles. The Mother Lode is by no means a single vein, but rather a system of linked veins, placed, however, within a narrow belt about a mile wide, and maintaining a remarkably straight course; it cuts Paleozoic slates and greenstones, but on the whole follows fairly closely a narrow belt of the Jurassic Mariposa slate and in places lies between this slate and the greenstone. There is a notable displacement along the Mother Lode, in the nature of a reverse fault (Fig. 189).

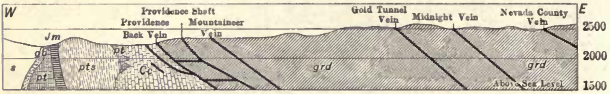

FIG. 190.—Geological section at Nevada City, California. *Cc*, Carboniferous slate; *Jm*, Jurassic slate; *pt*, porphyrite; *gb*, gabbro; *pts*, amphibolite; *s*, serpentine; *grd*, granodiorite. Scale 1 inch = 2,400 feet.

The strike of the veins is predominantly north-northwest, parallel to the range and to the strike of the steeply inclined strata, but the dip usually intersects that of the beds and, in the Mother Lode, is about 60° east. In many districts other directions of strike and dip prevail. The veins are easily traceable by prominent quartz outcrops, and many of them are remarkably straight and continuous in strike and dip. It is not uncommon to find veins continuous along the strike for 1 or 2 miles.

Many of the veins have been successfully worked to a vertical depth of 2,000 feet. In the Kennedy mine, on the Mother Lode belt, a vertical depth of 4,000 feet has been attained, good ore appearing in the lower levels. The Argonaut is now developed by a 4,500-foot incline reaching 4,000 feet in vertical depth. At the

Central Eureka mine, near the Kennedy, rich ore-bodies were found below the 1,000-foot level, though there was very little ore above that horizon. The North Star vein at Grass Valley has been followed along its flat dip for over 6,350 feet and, at that depth, yields a high production.

There are many structural types; the most common is the simple filled vein (Fig. 191), which may carry from a mere film of quartz to a thickness of 10 or 20 feet. Many of the outcrops appear to be much thicker than the figures just given, but these large masses are poor in gold. Again, there are composite veins or lodes in which certain belts of country rock are filled by branching veinlets of quartz or which may contain altered slabs of country rock (Fig. 193). In crushed clay slates the veins are sometimes broken and folded (Fig. 192). Large bodies of rock changed by replacement to gold ores are comparatively rare; such ores are mined in several places at Angels Camp, Calaveras County, but even here the gold is mainly contained in thin quartz seams in the altered rock. Again, gold-bearing quartz seams may follow joints of certain direction in large masses of rock; many such masses have been worked by the simple process

Fig. 191.—Argonaut vein in slate country rock, Amador County, California, at 650-foot level. *Photograph by O. H. Packer.*

of hydraulic washing of the upper, weathered part. Such deposits are called seam-diggings.

More rarely the veins follow narrow dikes of albite aplite; or they are developed on joint planes across the strike of thicker dikes in the manner of ladder veins.

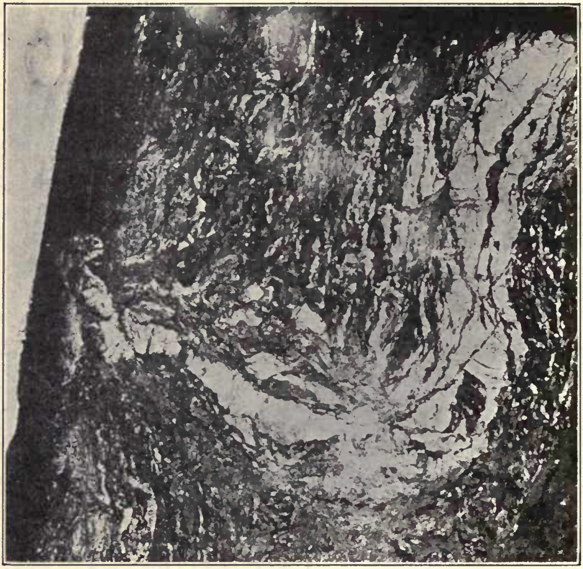

FIG. 192.—Bunker Hill vein, Amador County, California, showing folded vein in crushed clay slate.

The association of gold with dikes consisting mainly of albite rock has been described by Turner[1] and Reid.[2] Turner describes such dikes on Moccasin Creek, in Tuolumne County,[3]

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[1] H. W. Turner, Notes on the gold ores of California, *Am. Jour. Sci.*, 3d ser., vol. 47, 1894; *idem*, 3d ser., vol. 49, 1905; replacement deposits in the Sierra Nevada, *Jour. Geol.*, vol. 7, 1899, pp. 389–400.

[2] John A. Reid, The east country of the Mother Lode, *Min. and Sci. Press*, March 2, 1907.

[3] H. W. Turner, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, pt. 1, 1896, p. 664.

at the Shaw mine, in Eldorado County, and at other places, but the associated ores are generally of low grade and the mineralization is everywhere later than the dike.

Reid observed numerous thin dikes in Calaveras slate near Blue Canyon, Placer County, consisting largely of albite, which are cut or followed by seams or veins containing pyrite or arsenopyrite and native gold. The gangue in these veins consists of quartz and albite, with some manganiferous ankerite.

Along the walls there is always—in feldspathic and femic rocks adapted to such processes—more or less replacement extending a few inches or a few feet from the vein; bleaching, softening, and dissemination of pyrite indicate such replacement, in which the principal feature is the development of calcium-magnesium-iron carbonates with much sericite (p. 552). Occasionally, in sodic amphibolites, much albite develops, and in the vicinity of Angels Camp, on the Mother Lode, such replacements may contain enough gold to be called ore. In serpentine the alteration to a coarse aggregate of ankerite and bright-green chromium mica (mariposite) is characteristic; this product of replacement constitutes ore in only a few places, such as the Rawhide mine, southeast of Angels Camp, where it was penetrated by gold-bearing quartz stringers.

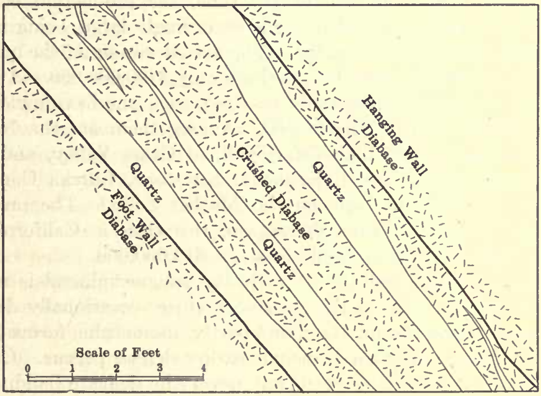

FIG. 193.—Section of Pittsburgh vein, Nevada City, California.

The ore-shoots are irregularly distributed; many veins are of pockety character, containing rich bonanzas at certain points, which may be determined by intersections or by the crossing of certain beds of the schist series. Other veins have large and regular shoots generally with a steep pitch, and sometimes with a pitch length of 2,000 or 3,000 feet. In isolated cases, such as the Idaho-Maryland vein at Grass Valley, the pitch of the rich pay-shoot was flat on the plane of the vein (p. 185). In many districts, especially at Grass Valley, the rule is that the shoot pitches to the left of an observer looking down the dip.

Including the placer gold yielded by the outcrops disintegrated during Tertiary and Quaternary time, the production of the California gold-quartz deposits is exceedingly large, being more than $1,200,000,000 in value. The actual mining of the quartz veins has yielded much less, perhaps only $400,000,000. A long list of celebrated mines could be cited, each one having yielded from $5,000,000 to $20,000,000. Among them are the North Star, Empire, and Idaho-Maryland, of Grass Valley, and the Plymouth Consolidated, Kennedy, Keystone, Eureka Consolidated, Gover, and Zeile on the Mother Lode.[1] The present annual production from deposits of this class in California is increasing and now amounts to about $13,000,000.

The principal and almost exclusive gangue mineral is milk-white quartz with coarse massive texture, occasionally drusy. In thin section the quartz shows partly idiomorphic forms (Fig. 48), and some individuals include earlier slender prisms. Comb structure is sometimes seen, but never the delicate banding of the veins formed near the surface. In places sulphides encrust rock fragments enclosed in quartz. A rough banding may result from irregular distribution of the sulphides, from the inclusion of narrow strips of black slate, or from subsequent shearing of the vein (Fig. 49); the last is not uncommon and is indicated in thin section by the crushing of the primary individual crystals (Fig. 54). Fluid inclusions are plentiful, and seem to consist of an aqueous solution. Carbon dioxide has been reported in one or two cases. Calcite, dolomite, and ankerite are formed in subordinate quantities, though they may be present abundantly in the replaced country rock adjoining the vein. Barite, fluorite, and tourmaline are practically absent,

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[1] The Kennedy mine to the close of 1915 has yielded $6,378,000 from 792,000 tons of ore. The North Star mine from 1884 to 1915 inclusive has yielded $17,450,526 from 1,358,394 tons of ore.

as are biotite, garnet, amphibole, epidote, zeolites, rhodonite, and rhodochrosite. No bituminous material has been reported. Mariposite, a chromium mica, is common near serpentine in the altered rock; roscoelite, a vanadium mica, is sometimes associated with native gold. Rutile is generally confined to the altered rock. Specularite and magnetite are absent, except in isolated cases. Scheelite is known to occur at several places.

The native gold is the principal ore mineral and occurs in all ores and at all depths. Sometimes large masses are found. A mass of solid gold valued at $40,000 was taken out from the Bonanza mine, near Sonora, in a pocket which yielded $360,000. This mine produced more than $2,000,000 in gold, the greater part of which was pounded out of the quartz in hand mortars. Still heavier masses of gold were found in the Monumental mine, Sierra County, and below the croppings of the Carson Hill veins on the Mother Lode. In some veins the gold is distributed in microscopic particles; in others it is visible (Fig. 194) and occurs in threads and plates. Coarse gold replacing quartz and arsenopyrite is described by Ferguson[1] from the pocket mines of Allegany, in Sierra County (Fig. 194, *C*). Very rarely, in some pocket mines, gold of a fineness exceeding 0.900[2] is encountered, but the average fineness is 0.800, and it is rarely as low as 0.700, the remainder *per mille* being principally silver.

Variable but always comparatively small quantities of metallic minerals accompany the gold, ordinarily making up 2 to 3 per cent. of the mass. Pyrite is universally present; pyrrhotite rarely, and then only in veins in granite rocks. Chalcopyrite, zinc blende, and galena are most abundant next to pyrite; arsenopyrite is not quite so common. Tetrahedrite is frequently found, while stibnite and molybdenite are rare. Compounds of tin, uranium, boron, phosphorus, and fluorine are lacking. Tellurides like altaite, hessite, calaverite, petzite, and melonite are sometimes associated with native gold.

The sulphides obtained by concentration from the ore are usually rich, often having a value of $100 to $300 per ton, but their value is only a small part of the value of the ore. C. G. Yale[3] gives the following figures for the Mother Lode mines. In the five Mother Lode counties 1,393,788 tons of ore were

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[1] *Bull.* 580, U. S. Geol. Survey, 1915, pp. 153–182.

[2] Gold from the San Giuseppe mine, near Sonora, was 0.990 fine.

[3] *Mineral Resources*, U. S. Geol. Survey, pt. 1, 1916, p. 217.

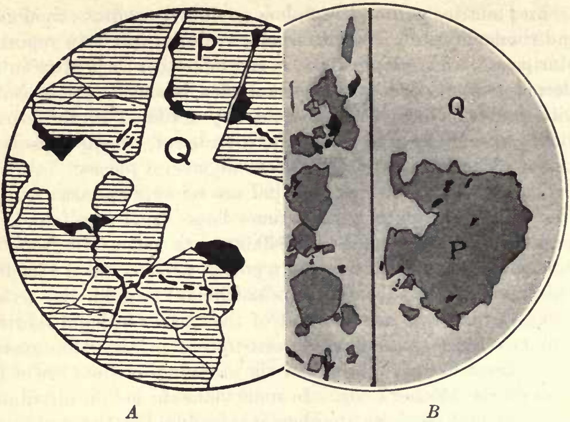
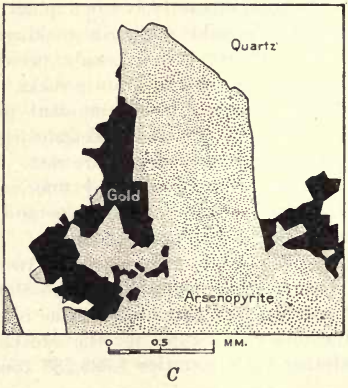


FIG. 194.—*A*, Thin section of gold-bearing quartz, Keltz mine, Tuolumne County, California. *Q*, quartz; *P*, pyrite; black, gold, deposited later than pyrite. Magnified 70 diameters. *After W. J. Sharwood.*

B, Thin section of gold-bearing quartz (*Q*), Omaha mine, Grass Valley, California, showing gold (black) deposited contemporaneously with pyrite (*P*). Magnified 17 diameters.



C, Gold replacing arsenopyrite and quartz, Alleghany district, California. *After H. G. Ferguson.*

mined in 1916, with a total recovery of $5,853,618. The gold recovery on the amalgamating plates or from the cyanide bullion averaged $2.69 per ton, while 36,974 tons of concentrates (mainly pyrite) obtained from the ore averaged in value $57 per ton; the total value recovered in gold (with a very small quantity of silver) averaged $4.20 per ton.

The ore in the large mines of the Mother Lode thus averages almost $4 per ton, which would be considered a very low grade of ore in districts like Tonopah, Cripple Creek or Pachuca. In some districts, like Grass Valley, where narrower veins are mined, the ore assays $10 to $15 per ton.

The gold-quartz veins of the Sierra Nevada were formed shortly after the intrusion of the granodiorite batholith in latest Jurassic or earliest Cretaceous time. They have, with the surrounding rocks, been subjected to an intense erosion, the vertical measure of which amounts to several thousand feet. The exposures by unequal erosion or by mining operations show, in many districts, that the vertical range of gold deposition without notable change in richness of shoots was over 4,000 feet; the relations in some districts lead to the conclusion that the deepest parts now mined were formed 7,000 feet or more below the surface.

The permeation of the metamorphic series by gold-bearing quartz is remarkable, although the greatest richness is concentrated, as stated above, in certain districts or along certain lines.

No hypothesis of lateral secretion can account for the great masses of quartz, nor for the occurrence of the veins in the most diverse rocks. For an explanation of their origin we are compelled to look to the great batholithic intrusion, or rather to the many minor intrusions on the flank of the range. At Grass Valley this conclusion is inevitable; on the Mother Lode it is less positive. The Mother Lode is, however, a profound dislocation, and we may well assume that it extends to a great depth and probably derived its metallic contents from underlying intrusive bodies. It must also be conceded that in many places the evidence points to the gabbros and peridotites (from which the serpentine was derived) and to the numerous albite aplite dikes which accompanied the basic intrusions as a source of at least part of the gold. A remarkable feature, nevertheless, is the absence, in the veins, of the usual mineralizers like chlorine, fluorine, and boron.

Regarding the nature of the depositing solutions forming gold-quartz veins we have, of course, no direct information. They must have been aqueous, to produce crustification of quartz and calcite. They must further have been competent to cause replacement by pyrite, sericite, and earthy carbonates. Hot waters containing carbon dioxide, alkaline carbonates, and hydrogen sulphide would fulfil these requirements. They probably carried gold dissolved in alkaline sulphides, a form in which the gold solution is stable to ordinary reducing agents such as carbon and pyrite. Such solutions will deposit the gold by contact with acids or by exposure to oxidation, probably also by decrease of temperature. W. Skey, T. Egleston. G. F. Becker, and lately V. Lenher[1] have drawn attention to this solvent.

The Gold-Quartz Veins of the Interior Cordilleran Region.[2]—A great number of intrusive masses of quartz monzonitic or granodioritic type are found in the interior Cordilleran region of the United States. They are, as a rule, of more recent age than the great coast batholith, their epoch of intrusion falling at the end of the Cretaceous or the beginning of the Tertiary. In or around these intrusives gold-quartz veins are often found, clearly related to the California type, but differing from it in some respects. Frequently they follow lamprophyric dikes (Fig. 195). They contain more sulphides, though of the same kinds, and they

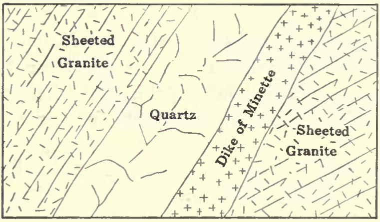


FIG. 195.—Section of Hidden Treasure vein, Neal district, Idaho.

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[1] *Econ. Geol.*, vol. 7, 1912, pp. 744–750; vol. 13, 1918, pp. 161–184.

[2] W. Lindgren, The mining districts of Idaho Basin, etc., *Eighteenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1897, pp. 617–744.

W. Lindgren, The gold belt of the Blue Mountains of Oregon, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, pp. 551–776.

C. E. Weaver, The Blewett mining district, *Bull.* 6, Washington Geol. Survey, 1911.

carry, as a rule, more silver in the sulphides than the veins of the California type; there is less free gold, and, in some instances, only a small proportion of the total gold is amenable to direct amalgamation. The gold contains silver and rarely has a fineness above 0.700. Rich silver minerals often form in the oxidized zones. The precious metals are contained chiefly in the quartz filling, but the altered rock adjoining the veins sometimes carries gold and silver, which means that it is in part replaced by gold- and silver-bearing sulphides. In feldspathic and ferromagnesian rocks sericitic and pyritic alteration (Fig. 196) affects the wall rocks; carbonatization is often observed, but is rarely as intense as in the veins of the Sierra Nevada. Pyrite, arsenopyrite, chalcopyrite, galena, and blende are the common ore minerals, but tetrahedrite is also plentiful and cinnabar is known to occur. Tellurides are sometimes present and are almost intergrown with native gold. Quartz is the prevailing and usually the only gangue mineral. Quartz with coarse native gold has been observed to replace an earlier calcite gangue (Fig. 197). Tourmaline, magnetite, and pyrrhotite are not known. The grade of the ore is from $5 to $15 per ton.

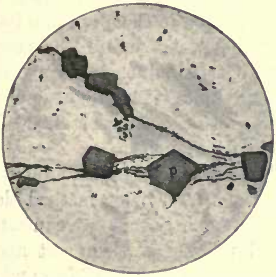

FIG. 196.

FIG. 197.

FIG. 196.—Pyrite (*p*), forming by replacement along calcite veinlets (black); calcite forms lining around pyrite crystals. In chloritic diabase, Great Northern mine, Canyon, Oregon. Magnified 10 diameters.

FIG. 197.—Quartz (*q*) with native gold (black), replacing vein of calcite (*c*); Great Northern mine, Canyon, Oregon. Magnified 10 diameters.

Victoria, Australia.[1]—The principal gold-bearing region of Victoria, though of much smaller extent than the California gold belt, is believed to have produced about the same amount, namely, $1,300,000,000 in gold. Here, too, the placers have yielded by far the greater production. Both gravel deposits and quartz veins still yield a gradually diminishing output. In 1915 the production of gold from quartz mines was only about 218,660 ounces. The ores averaged $7.50 per ton.

This most productive region includes the celebrated districts of Ballarat and Bendigo and is situated in the low ranges of the mountains rising between the basaltic and Tertiary terranes on the south and the Murray Plains on the north (Fig. 198).

Little-altered Ordovician slates and sandstones prevail and form sharply compressed folds. Intruded in them are two batholiths of granitic rock, probably quartz monzonite, the largest being that between Bendigo and Castlemaine; there are also many smaller bodies of the same kind. The intrusions are probably of late Silurian age, and erosion of at least 3,000 feet has planed the region to an undulating surface.

Within the folded Ordovician rocks quartz veins are abundant and generally follow the strike of the strata, being massed along certain productive "reef lines." Frequently they are conformable between shale and sandstone, but some of them cut across the strike. A common type has one well-defined wall from which flat and irregular bodies of quartz project into the hanging or foot wall. These flat "makes" are particularly characteristic and usually contain the best pay at Ballarat East and other places. The saddle reefs constitute an interesting division, in which masses of quartz fill cavities produced at anticlines (Fig.

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[1] E. J. Dunn, Report on the Bendigo gold field, Dept. of Mines, Melbourne, 1896.

T. A. Rickard, The Bendigo gold field, *Trans.* Am. Inst. Min. Eng., vol. 20, 1891, pp. 463–545.

J. W. Gregory, The Ballarat East gold field, *Mem.* 4, Victoria Geol. Survey, 1907.

W. Baragwanath, The Castlemaine gold field, *Mem.* 2, Victoria Geol. Survey, 1903.

O. A. L. Whitelaw, The Wedderburn gold field, *Mem.* 10, Victoria Geol. Survey, 1911.

W. Lindgren, Characteristics of gold-quartz veins in Victoria, *Eng. and Min. Jour.*, March 9, 1905.

F. L. Stillwell, Replacement in the Bendigo quartz veins, etc., *Econ. Geol.*, vol. 13, 1918, pp. 100–111.

199) or less commonly at synclines (Fig. 200) by stresses subsequent to the principal folding; they are often connected with irregular "makes" and spurs (Fig. 201) of quartz. These open cavities, subsequently filled by quartz, are the necessary result

Fig. 198.—Sketch map of eastern Australia, showing location of important mining districts.

of stresses applied to folded masses of little-altered sediments, the strata of which vary considerably in hardness.

The best instances of saddle reefs, many of them superimposed upon and following three or four distinct lines of anticlines, are found at Bendigo (Fig. 36) and Castlemaine. The Bendigo veins have been worked to a depth of 4,600 feet in the Victoria reef, situated on the New Chum reef line, but sinking has been suspended. A body of quartz, containing at best $17 per ton, was mined at a depth of about 4,200 feet, but it is said that on the whole little profitable mining has been done at Bendigo below a depth of 2,500 feet. The granite rocks rarely contain quartz veins. The vein-filling is a massive milk-white, sometimes glassy quartz of coarse crystalline texture. It contains native gold, often coarse, and also a little pyrite and arsenopyrite; sometimes also a little galena, zinc blende, molybdenite, and stibnite. No tellurides are reported. There is neither barite nor fluorite. Ankerite or calcite with some magnesium and iron is common but subordinate, usually appearing near the walls. Albite and a vermicular chlorite are present in places, the former in vugs, the

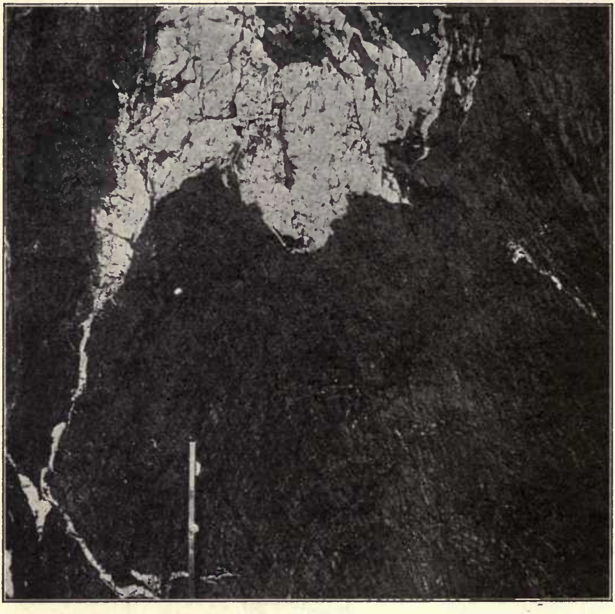

FIG. 199.—Saddle reef in slate and sandstone, Bendigo, Victoria. *After E. J. Dunn.*

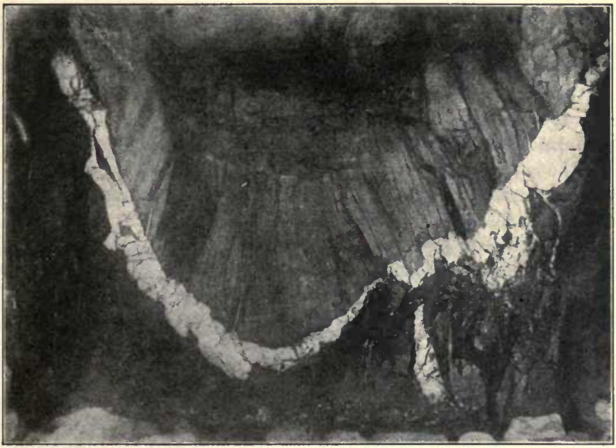

FIG. 200.—Trough reef, in slate and sandstone, Bendigo, Victoria.
After E. J. Dunn.

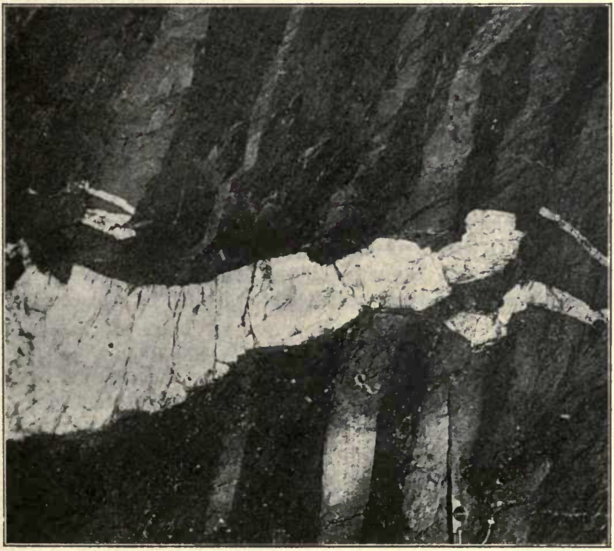

FIG. 201.—Spur reef in slate and sandstone, Bendigo, Victoria.
After E. J. Dunn.

latter enclosed in massive quartz. There is little evidence of banded structure, except that near the walls of the veins thin lamellæ of slate may be interlaminated with quartz.

At Ballarat rich ore-bodies occur at the intersection of flat bodies of quartz with certain thin pyritic and carbonaceous seams of slate, the so-called "indicators." It has been held that the gold has been precipitated by the carbon in the indicator. A more plausible view is that the indicators are narrow fissures, later than the flat "makes" and enriching them at the intersection. Similar features have been noted at other points in Victoria and seem to point to a process of enrichment, although probably not caused by surface waters. At Ballarat the developments at depths below 1,500 feet have not been encouraging.

Australian geologists have presented strong evidence that the deposition of the quartz was completed before the Devonian rocks were laid down, and this determines the age of the veins within narrow limits. The granitic intrusion and the formation of the quartz veins were closely associated events. The fact that so few lodes occur in the granitic rocks is probably explained by the great resistance of the hard intrusive bosses to compressive stresses, compared with the yielding nature of the soft sedimentary rocks.

J. R. Don (p. 13) has shown that the sediments away from the veins contain no gold, and that the increasing traces of gold found as the veins are approached are dependent upon the amount of pyrite introduced from the veins.

Metasomatic processes play but a small part in these veins. The slates are little altered, except by the introduction of pyrite and occasionally of some carbonates of calcium and magnesium. Dunn held that the veins had been opened by the crystallizing force of the quartz. Recently Stillwell has advanced the view that the laminated "ribbon quartz" has been formed by replacement of shale.

New South Wales and Queensland.—A large number of well-known and productive districts are found in New South Wales and Queensland, in which the gold occurs in quartz veins associated with intrusive rocks. Some of these veins carry quantities of sulphides besides free gold; occasionally fluorite and barite are reported. The almost universal conditions are a deeply eroded region with diorite or granodiorite or their basic dikes intruded into Paleozoic sediments which are usually more highly altered than in Victoria; the veins occur either in intrusive or in sedimentary rocks or in both. Placers are usually present.

At Hill End,[1] north of Bathurst in New South Wales, folded Silurian slates and tuffs are intruded by dikes and sills of quartz porphyry. The lenticular quartz veins lie in slate or at the contact with the intrusive rocks. Coarse gold prevails; one mass extracted in 1872 consisting of solid gold mixed with some quartz weighed 630 pounds and was valued at $60,000. Five and one-half tons of solid gold were recovered at this place from 10 tons of quartz, the value of the gold being $3,300,000. Similar geological conditions exist at Hargraves, but the quartz here occurs as saddle reefs. Here, as at Ballarat, flat "makes" are present and are enriched where they are crossed by "indicators" or narrow bands of dark-greenish slate.

At Hill Grove,[2] in the New England district, in the northeastern part of New South Wales, near the wolframite deposits mentioned elsewhere (p. 671), slates and quartzites are intruded by quartz-mica diorite and the veins occur in the sedimentary rocks, often near lamprophyric dikes. The veins, which average 6 feet in width, contain, besides quartz and free gold, scheelite, arsenopyrite, and much stibnite. Andrews regards them as due to the last emanations from the same granitic magma, the earlier high-temperature emanations having produced the cassiterite-molybdenite-wolframite deposits.

At Charters Towers,[3] in Queensland, the veins intersect granitic rocks ranging from granites to tonalites or quartz-mica diorites. The veins have been highly productive and have been worked to a depth of 3,000 feet along the dip; they contain about 7 per cent. of sulphides (pyrite, galena, zinc blende, pyrrhotite, and arsenopyrite). As usual in granitic rocks, a considerable part of the gold of the ore is contained in the sulphides. The veins are regular but narrow, averaging about 3 feet in thickness. The average value of the ore is probably less than $15 per ton.

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[1] E. F. Pittman, Mineral resources of New South Wales, Geol. Survey, N. S. W., 1901, p. 31. Comprehensive summary in Maclaren's "Gold," 1908, pp. 341–358.

[2] E. C. Andrews, *Records*, Geol. Survey, N. S. W., vol. 8, 1909, p. 143.

[3] Jack, Rands, and Maitland, *Ann. Rept.*, Geol. Surv., Queensland, 1892. W. E. Cameron, *Publ.*, No. 224, Geol. Survey, Queensland, 1909.

Nova Scotia.—The gold-quartz veins of Nova Scotia,[1] from which during the last fifty years a moderate production has been derived, are, in many respects, of special interest. The veins are contained in folded sedimentary rocks—slate and quartzite—probably of Cambrian age, and these are intruded by granitic rocks of Silurian age. The gold belt extends for a distance of 280 miles along the south coast, and its average width is about 30 miles. The numerous quartz veins, many of which can be traced for long distances, often occur in the manner of saddle reefs along the anticlines. The anticlinal axes are in places marked by structural elliptical domes in which the strata pitch both ways on the strike, and gold-bearing quartz veins are usually found in such domes. The veins are ordinarily parallel

FIG. 202.—Folded quartz vein in slate, Mexican mine, Goldenville, Nova Scotia. *After T. A. Rickard.*

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[1] E. R. Faribault, The gold measures of Nova Scotia and deep mining, *Jour.*, Canadian Min. Inst., vol. 2, 1899, pp. 119–161.

J. E. Woodman, Geology of Moose River gold district, Nova Scotia, Inst. Nat. Sci., vol. 11, 1903, pp. 18–88.

W. Malcolm, Goldfields of Nova Scotia, *Mem.* 20-E, Canada Dept. Mines, Geol. Survey Branch, 1912.

to the stratification, but some of them, while parallel in strike, cut across the dip. Corrugated and crenulated veins are common (Fig. 202) and the term "barrel quartz" is used to describe the material in them; the corrugation is believed to have been caused by deformation subsequent to the deposition. The gangue is always quartz; arsenopyrite, pyrite, chalcopyrite, galena and zinc blende are fairly common, but the principal valuable mineral is native gold.

Veins with stibnite occur in the auriferous belt, and scheelite has been recently discovered.[1] Faribault holds that the intrusive granite, in which no gold-quartz veins have been found, is later than the veins. If this is the correct interpretation, the relationship is the reverse of that in all other districts containing similar veins. T. A. Rickard, however, has expressed a contrary opinion and believes that the formation of the gold-bearing veins succeeded the granitic intrusion.

Under the microscope the glassy quartz shows intense deformation and the corrugated veins are probably simply the result of the crumpling of harder beds in a plastic medium (cfr. Fig. 14). There were probably two epochs of folding, one preceding and the other following the deposition of quartz. T. A. Rickard,[2] on the other hand, believes that the crenulations are the product of complex fractures in rocks of uneven texture.

GOLD-ARSENOPYRITE DEPOSITS

The frequent association of arsenopyrite with gold has been noted above.

In some veins arsenopyrite predominates and the ore may be utilized for the recovery of arsenic as a by-product. Some of these deposits contain also pyrite, chalcopyrite, pyrrhotite, zinc blende, galena, realgar and stibnite in the order of succession named. Quartz is the prevailing gangue mineral. Arsenopyrite is always the oldest ore mineral.[3]

Similar ores are also found as lenses and replacement veins in schist and then usually show affiliations with the high tempera-

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[1] V. G. Hills, Tungsten mining in Nova Scotia, *Proc.* Colorado Sci. Soc., vol. 10, 1912, pp. 203–210.

[2] The domes of Nova Scotia, *Trans.* Inst. Min. and Met., London, vol. 21, 1912, pp. 506–560.

[3] J. E. Spurr, The ore deposits of Monte Cristo, *Twenty-second Ann. Rept.*. U. S. Geol. Survey, pt. 2, 1901, pp. 777–865.

ture deposits. Many such deposits of arsenopyrite occur in Ontario,[1] for instance, at the Deloro Mine, which for long time was worked for gold and arsenic.

GOLD-BEARING REPLACEMENT DEPOSITS IN LIMESTONE

Deposits in which limestone is replaced by jasperoid or fine-grained silica and which carry gold or silver or both are sometimes found in the Cordilleran States where intrusive porphyries invade calcareous sediments. Few examples are known elsewhere. These ores, which are usually very poor in sulphides, are at several places of great economic importance.

In the Mercur district,[2] situated in the Oquirrh Range in Utah, such siliceous silver ores are found at the lower contact of a thin sheet of granite porphyry with Carboniferous limestone. The jasperoid rock, in places 55 feet thick, contains more or less silver throughout but has not been extensively worked. It carries barite and calcite in places and large cavities are sometimes covered by crystals of these minerals. The ore contains some stibnite, also a little copper and very small amounts of arsenic, molybdenum, and tellurium. No pyrite was observed.

In the same district, below an upper sheet of porphyry which like the lower is greatly decomposed by processes of weathering, is found a sheet of jasperoid rock locally 25 feet thick which contains minutely divided, generally invisible gold with some fine-grained pyrite, a little barite, and some realgar and cinnabar. In part the porphyry itself constitutes ore and the ore may extend into the limestone above the porphyry. Spurr suggests that the ores gained access to the sheet through vertical fissures, now filled with calcite. The deposits are evidently later than the porphyry intrusion, and the ores are similar to those of the Black Hills.

From 1890 to the end of 1913 about 4,900,000 tons of this gold ore averaging about $3.58 per ton in gold have been mined in the Mercur district.[3] The total yield having had a value of $19,000,000. The mines are now closed and dismantled.

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[1] W. G. Miller and C. W. Knight, Ontario Bur. Mines, vol. 22, pt. 2, 1914, p. 110; also *idem*, vol. 11, 1902, p. 105.

[2] R. C. Hills, *Proc.*, Colorado Sci. Soc., Aug. 6, 1894.

J. E. Spurr, *Sixteenth Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1895, pp. 349–455.

[3] V. C. Heikes, *Mineral Resources*, U. S. Geol. Survey, 1913 and previous years, Production of gold and silver, chapter on Utah.

W. H. Weed[1] describes similar deposits in the Moccasin district, in Montana, where rhyolite porphyry and phonolitic dikes intrude Carboniferous limestone. Near these intrusives the limestone is replaced by fluorite (Fig. 64) and by jasperoid; the replaced rock contains gold and has been mined successfully for a number of years. Some of the ore deposits here, too, lie at the lower contacts of intrusive sheets.

The so-called refractory siliceous ores of the Black Hills of South Dakota, described by J. D. Irving,[2] constitute the best examples of this type of replacement ores. For many years these ores have yielded annually about $2,000,000 in gold and 100,000 ounces of silver, from about 600,000 short tons. The ores are treated by the cyanide process. The deposits form replacements of dolomite at two horizons in the Cambrian section of the Black Hills, in a region which is intruded on a large scale by dikes, sheets, and laccoliths of rhyolite porphyry, syenite porphyry, and phonolite of probable Eocene age.[3] The more important lower horizon is 15 to 25 feet above the basement of pre-Cambrian schists.

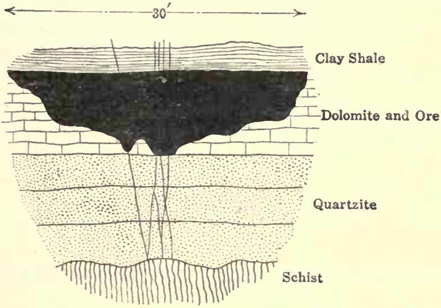

Fig. 203.—Cross section of shoot of siliceous ore replacing Cambrian dolomite, Black Hills, South Dakota. Spread of ascending solutions on under side of impervious shale makes shoot wider at top. *After J. D. Irving.*

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[1] W. H. Weed and L. V. Pirsson, Geology and mineral resources of the Judith Mountains of Montana, *Eighteenth Ann. Report*, U. S. Geol. Survey, pt. 3, 1898, pp. 437–616.

[2] *Prof. Paper* 26, U. S. Geol. Survey, 1904.

[3] T. A. Jaggar, *Idem*, pp. 24–26.

The ores occur immediately below more or less impervious beds of shale or below sills of intrusive rocks. While the richest ore replaces the dolomite, ores of lower grade may also replace the underlying basal Cambrian quartzite and the overlying shale; the replaced bodies are at most 18 feet thick, averaging 6 feet. These channel-like ore-bodies have a width attaining 300 feet but averaging much less. Their length is considerable, one shoot being followed for three-fourths of a mile; many parallel shoots may be found in one locality, each shoot corresponding to a fissure or series of fissures (vertical) which intersect the basal beds but which do not carry the ore below the quartzite and rarely above the shale (Fig. 203).

The ore is a hard, brittle fine-grained siliceous rock, often reproducing the dolomite texture with great fidelity (Fig. 56). The fresh ore is locally bluish and contains finely divided pyrite; much of it contains solution-cavities lined with quartz crystals. Fluorite is always present, frequently also barite. Other associated minerals are stibnite, occasionally wolframite, and probably arsenopyrite and tellurides in fine distribution. Much of the ore is mined at shallow depths and it is largely oxidized. Interesting data as to the form and distribution of the ore-shoots are also given by J. D. Irving in a later paper.[1]

GOLD-BEARING REPLACEMENT DEPOSITS IN QUARTZITE

It is not uncommon in mineralized districts to find gold ores in quartzite. Usually they take the form of gold-quartz veins, but replacement ores may also occur. Such a deposit is that of the Delamar mine,[2] in central Nevada, which for many years had a large production, aggregating several million dollars in gold, but which is now closed. At this place Paleozoic quartzite is intruded by two dikes of granite porphyry and one narrow lamprophyric dike. Along the latter a strong fracture has developed. The ore occurred in an irregular chimney on both sides of this fracture and is later than the dikes (Fig. 204). The ore has been formed by recrystallization of the quartzite and does not appear in the dikes. The rock of granular texture is replaced by a fine-grained drusy aggregate, with a little pyrite

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[1] Replacement ore-bodies, *Econ. Geol.*, vol. 6, 1911, pp. 527–561.

[2] S. F. Emmons, *Trans.*, Am. Inst. Min. Eng., vol. 31, 1901, pp. 658–675.

and a telluride of gold. Only the oxidized ore was worked; the tenor of this gradually decreased until on the tenth level it was only $4 to $5 per ton and below working cost.

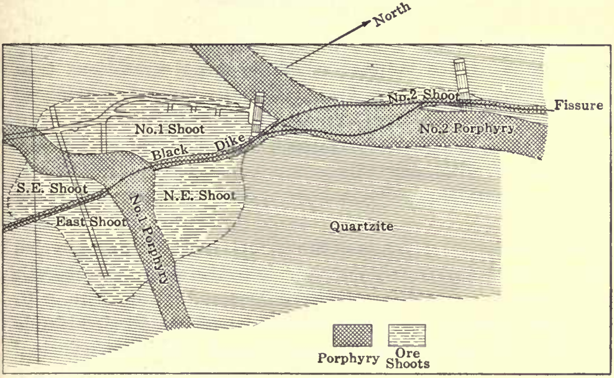

FIG. 204.—Plan of seventh level of Delamar mine, Nevada. *After S. F. Emmons.*

GOLD-BEARING REPLACEMENT DEPOSITS IN PORPHYRY

Replacement deposits in aluminous rocks like granite porphyry, gneiss, and amphibolite are not uncommon in connection with gold-bearing veins, as shown, for instance, at Cripple Creek, Colorado, and along the Mother Lode in California (pp. 523 and 571). Larger bodies of rock are more rarely replaced. W. H Emmons[1] has described an example of this in the Little Rocky Mountains of northeast Montana, a small outlier on the Great Plains. Stocks and sheets of syenite porphyry are intruded in a Paleozoic sedimentary complex. Broad zones in this porphyry are replaced and cemented by quartz, pyrite, secondary orthoclase, and fluorite. The deposits are really wide replacement lodes, some of them traceable for 1,200 feet and varying from a few feet to 100 feet in width. The gold is finely distributed and probably occurs as a telluride, the ores averaging about $3 per ton in gold and one ounce of silver. The operations have thus far

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[1] *Bull.* 340, U. S. Geol. Survey, 1908, pp. 96–116. See also W. H. Weed and L. V. Pirsson, *Jour. Geol.*, vol. 4, 1896, pp. 399–428.

been confined to the oxidized zone, which descends to a depth of 200 feet. The relationship to the Cripple Creek ores is evident.

THE SILVER-LEAD VEINS

General Features.—The silver deposits of intermediate depths include many types between which so many transitions exist that a classification is difficult. Certain forms occurring as fissure veins parallel closely the gold-bearing quartz veins; but many of the silver deposits contrast with those of gold in being associated with carbonate gangue, more frequently ankerite or other magnesium-calcium-iron carbonates than calcite or siderite.

The replacement deposits in limestone very often contain rich silver ores, though rarely much gold. The two ore minerals most common in silver deposits are galena and tetrahedrite; with these zinc blende is usually associated. Galena and zinc blende may so predominate that the base metals yield the principal value of the deposits. Chalcopyrite, pyrite, and arsenopyrite play subordinate parts. Native silver is probably never a primary mineral, although abundantly formed by secondary reactions effected by descending waters, and rich sulphantimonides like proustite, pyrargyrite, and polybasite are also largely, though not wholly, of similar secondary origin.

The following types merely serve as centers around which the descriptions may be grouped.

Quartz-Tetrahedrite-Galena Veins.—Prominent veins carrying milky quartz and sparsely disseminated tetrahedrite, galena, and zinc blende, with subordinate pyrite, are common in the Cordilleran region in or near intrusive bodies of granitic texture. Many such veins are found in the great batholiths of Idaho and Montana; also in New Mexico—for instance, at Organ,[1] where quartz monzonite breaks through Paleozoic limestones. The deposits are on the whole poor and rarely worked, although from 1870 to 1890 the enriched surface zones in many places yielded much silver chloride, native silver, and ruby silver. The Granite-Bimetallic vein in Montana is a famous representative (p. 888). With rising silver prices many such veins will be re-opened.

Tetrahedrite-Galena-Siderite Veins (Wood River Type).—The association of siderite gangue with galena and zinc blende and

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[1] *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 209.

with a smaller quantity of tetrahedrite rich in silver (freibergite) is not uncommon in veins associated with intrusions of quartz monzonite, granodiorite, diorite, or lamprophyric dikes. The latter dikes are the latest igneous rocks, and the ores appear to have been introduced shortly after their intrusion.

The deposits are usually veins in which the ores appear in part as filling, but largely as replacements of the country rock. Siderite is the characteristic gangue mineral, but calcite and intermediate carbonates of calcium, magnesium, and iron are often present; quartz enters into the gangue when the veins intersect the granitic rocks. Among the ore minerals tetrahedrite is the principal carrier of silver and is often intimately intergrown with galena. The galena is mostly coarse grained and also carries silver, while the zinc blende, with 4 or 5 per cent. of iron, is relatively poor in silver, but is sometimes recovered as a by-product in concentration.

Chalcopyrite is less abundant than tetrahedrite; pyrite is not conspicuous; while arsenopyrite and pyrrhotite occasionally appear, particularly in granitic country rock.

The ore-bodies often replace calcareous shales along the vein, but these shales appear to have been little altered, except for the introduction of metallic minerals and some siderite. In granular, feldspathic rocks, close to the vein, sericite, carbonates, and a chlorite rich in iron develop in large amounts and sulphides are introduced. A complete replacement by sulphides is unusual. Sodium is almost wholly removed, but potassium fixed as sericite and calcium fixed as carbonate remain.

The structure of the ore is generally massive, and large bodies of galena are common; in one stope of the Minnie Moore mine, in the Wood River district, Idaho, 16 feet of solid galena was shown. Smaller veins may show banded structure, but rarely comb structure. Sometimes a thin layer of quartz may be found along the wall, then a narrow comb of calcite, while the mass of the vein consists of massive galena, alternating with bands of zinc blende and containing intergrown tetrahedrite. The tetrahedrite appears to replace galena. The galena frequently shows phenomena of pressing, gliding and recrystallization.

The width of the veins rarely exceeds a few feet, and part of this is usually crushed country rock. Their outcrops are inconspicuous.

The ore-shoots are markedly irregular and the cost of mining is therefore high. A marked deterioration may often be observed in depth; the large bodies of rich silver ore—aside from those affected by the surface enrichment—are found comparatively near the surface and the lower levels commonly show pinched veins or a predominance of zinc blende, pyrite and quartz.

The upper, oxidized parts of the veins are usually enriched by secondary silver chloride, native silver, and pyrargyrite.

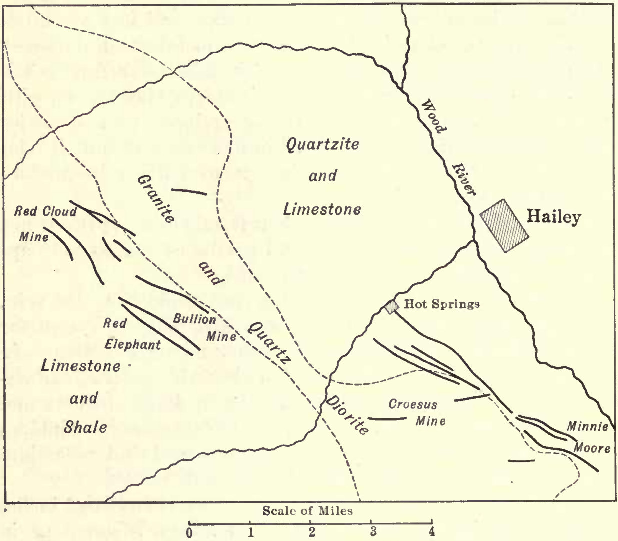

FIG. 205.—Vein systems near contact of intrusive mass in Carboniferous sediments, Wood River district, Idaho.

Wood River, Idaho.[1]—The silver-lead veins near Hailey, Idaho, on the Wood River, north of the Snake River lava plains, were discovered about 1878 and for many years yielded a high production which now has decreased greatly.

The district lies a few miles east of the eastern contact of the great granitic batholith of central Idaho and the prevailing rocks are calcareous shale, quartzite, and limestone of Carboniferous

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[1] W. Lindgren, Wood River mining district, *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, pp. 218–231.

age, compressed in northwestward-striking folds. These sedimentary rocks are intruded by a minor batholith of diorite and quartz monzonite, following the general direction of the strata and from 2 to 3 miles wide. The deposits are fissure veins arranged in two parallel linked systems (Fig. 205) along the contacts of the batholith and in places cutting across the contacts into the granitic rock. Some of the veins follow lamprophyric dikes. Their strike generally cuts the stratification at an acute angle and their dip is prevailing 50° southwest. The croppings are inconspicuous. Most of the veins are in calcareous shale.

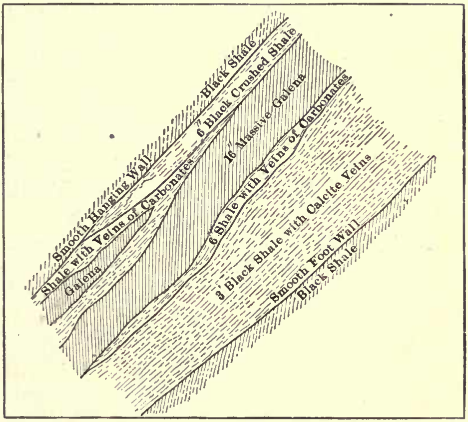

FIG. 206.—Section of vein in Wood River district, Idaho.

The ore consists of galena, zinc blende, and tetrahedrite, with but little pyrite and chalcopyrite; the gangue is siderite, or intermediate calcium, iron, and magnesium carbonates, with a little quartz. The ore minerals have massive structure, sometimes roughly banded. Second-class ore consists of seams of carbonate gangue with small grains of galena. As there are no smelting works in the district the ores and concentrates must be shipped; the shipments consist of high-grade ore containing 40 to 50 per cent. of lead and 50 ounces of silver ton; a little gold is usually present.

Most of the veins are narrow, although they may in places widen out into bodies of galena many feet wide. The typical structure of a filled vein is shown in Fig. 206.

The ore-bodies are irregularly scattered along the veins and are for the most part replacements of calcareous shale by galena. Some of these replacement bodies lie obliquely across the strike of the vein and may be several hundred feet long and 10 to 30 feet wide. In a few places the developments have been carried far below the adit levels, and, on the whole, the lowest levels have shown fewer and poorer ore-bodies than the upper parts of the veins. There is, however, little indication of sulphide enrichment, and the oxidized zone is shallow.

While some of the veins in the granitic rocks have the same character as those in the shales, others carry gold as the principal metal. These gold-bearing veins occur both in the diorite of the small batholith and in the main batholith of quartz monzonite; they contain quartz, calcite, siderite, pyrrhotite, arsenopyrite, and chalcopyrite and are in part free milling. Without doubt, these gold deposits belong to the same epoch of metallization as the silver-lead veins, and they show the same type of metasomatic alteration, namely, sericitization and carbonization.

Slocan, British Columbia.[1]—The veins of the Slocan district are mainly contained in the clay slates of the Slocan series, the age of which is possibly Carboniferous. The sedimentary rocks are intruded by granite, quartz porphyry, and lamprophyric dikes. The fissure veins have a general northeast direction and high southeast or northwest dips. Where the veins intersect the igneous rocks quartz is the prevailing gangue mineral. In the sedimentary rocks the gangue is mainly siderite or manganosiderite. A specimen gave, for instance, 59 per cent. $FeCO_3$, 27 per cent. $MnCO_3$, 12 per cent. $MgCO_3$, and 2 per cent. $CaCO_3$. The ore minerals are zinc blende, galena, and tetrahedrite rich in silver. Pyrite and chalcopyrite are fairly common; pyrrhotite is less abundant and is confined mainly to the vicinity of intrusive rocks. Native silver, of secondary origin, is present and sometimes coats the cleavage planes of zinc blende.

In the Slocan district also gold-bearing veins occur together with the silver-lead veins and are apparently of the same age.

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[1] O. E. Leroy, Summary report for 1909, Geol. Survey Canada, 1910, pp. 131–133; also Geol. Survey Canada, Map 62A, 1912.

W. R. Ingalls, Report of the Zinc Commission, Ottawa, 1906, p. 238, etc.

W. L. Uglow, *Econ. Geol.*, vol. 12, 1917, pp. 643–662.

Many of the Slocan veins have proved less rich in depth than near the surface and contain more siderite, quartz, and pyrite; this is probably a change in the primary mineralization. The succession of minerals is: Siderite, sphalerite, galena, tetrahedrite.

Galena-Siderite Veins.—The galena-siderite veins form a small but important type, represented in the United States in the Coeur d'Alene district[1] in Idaho. In contrast to the Wood River type these veins contain little tetrahedrite and are poor in silver; on the other hand, they yield about one-third of the lead production of the United States and in the aggregate also much silver. In 1916 the ore treated amounted to 2,500,000 short tons which yielded about 178,000 tons of lead and 11,600,000 ounces of silver; the estimated yield of zinc was 43,000 tons. Among the principal mines are the Bunker Hill & Sullivan, the Interstate-Callahan, the Hecla, the Morning, and the Hercules. Some of the mines have been worked to a depth of 3,300 feet below the outcrops.

The prevailing country rock is a fine-grained sericitic quartzite, referred to the Burke and Revett formations of the thick and closely folded pre-Cambrian Belt series of northern Idaho. These rocks are traversed by many faults, which, however, are not mineralized; the veins follow subordinate fissures of small throw. Two masses of monzonite of probable Cretaceous age, the larger not more than 3 miles in length, intrude the Belt series and cause some contact metamorphism by the development of biotite, garnet, and pyroxene in the quartzites. The last phases of the intrusion are represented by lamprophyric dikes. These are later than the mineralization and intersect the ore.

It is held probable that the intrusions of monzonite connect and widen below the surface. The ore deposits are composite veins or lodes, often of considerable thickness, formed partly by filling, but largely by replacement of the country rock along nearly vertical shear zones with northwesterly trend. The

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[1] F. L. Ransome and F. C. Calkins, The geology and ore deposits of the Coeur d'Alene district, Idaho, *Prof. Paper* 62, U. S. Geol. Survey, 1908.

J. R. Finlay, The mining industry of the Coeur d'Alenes, Idaho, *Trans.* Am. Inst. Min. Eng., vol. 33, 1903, pp. 235–271.

O. H. Hershey, Genesis of the silver-lead ores in the Wardner district, Idaho, *Min. and Sci. Press*, June 1, 8 and 15, 1912; also Sept. 17, Oct. 4, 1913; May 20, 1916.

V. C. Heikes, Mines report on Idaho in *Mineral Resources*, U. S. Geol. Survey, pt. 1, Annual publication.

longest of the veins is the Bunker Hill, which is traceable for 7,000 feet.

The ore-shoots are large, and many of them are roughly vertical; some have been followed for 3,000 feet in pitch length. At the Bunker Hill & Sullivan mine (Fig. 207) the ore-bodies do not always follow the main wall, which dips 38° SSW., but may lie in the shattered country rock within 250 feet above it. The width of the ore is in places as much as 40 feet, 9 feet being the average in some of the larger mines.

Galena, with some pyrite and zinc blende, and in places a little tetrahedrite rich in silver are the principal ore minerals. Chalcopyrite is present in small amounts; in some mines pyrrhotite takes the place of pyrite. Siderite and quartz are the predominant gangue minerals; barite, calcite, and dolomite are rare. Some of the siderite contains several per cent. of manganese.

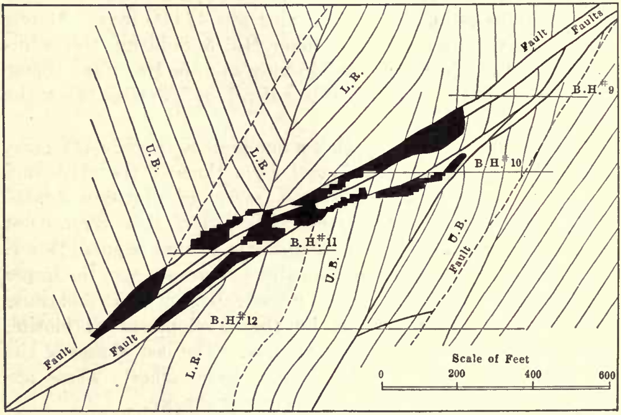

Fig. 207.—Vertical cross-section of part of Bunker Hill and Sullivan vein showing relation of ore-bodies (black) to structure. *L. R.*, Lower Revett formation; *U. B.*, Upper Burke formation. Lines between faults indicate intersections of stratification planes with vertical plane. *After Oscar H. Hershey.*

The ores are in large part formed by replacement of sericitic quartzite along the tight shear planes of the lodes. The siderite develops first, replacing both sericitic cement and quartz grains in the quartzite. Rhombohedrons of siderite may often be seen cutting across the clastic grains. According to Ransome the galena is in part later than the siderite and replaces that mineral. Replacement of quartzite by galena is shown in Fig. 60. Some ore-bodies consist of almost massive galena, but the ordinary ore is an aggregate of siderite and galena, which must be concentrated. The bulk of the ores range from 3 to 14 per cent. lead and from 2.5 to 6 ounces of silver to the ton. These are concentrated to a product containing about 50 per cent. of lead. The lowest grade which can be worked at present carries 5 per cent. lead and 3 ounces of silver to the ton. A few of the mines like the Interstate-Callahan and the Morning yield much zinc blende.

The metasomatic action indicated by the presence of siderite in the quartzite often spreads for 100 feet or more beyond the ore. There are some indications of change of ores in depth; Ransome finds that in the lower levels of many mines pyrite, pyrrhotite, and zinc blende become more abundant.

Ransome traces a genetic connection between the ore deposits and the intrusive monzonite. Contact-metamorphic ores in irregular bodies are found in two mines close to the monzonite, in the contact zone. These ores contain galena, zinc blende, pyrite, pyrrhotite, chalcopyrite, and magnetite, with a gangue of garnet, biotite, and diopside. In veins near the intrusive mass pyrrhotite and magnetite, as well as garnet and biotite, are found. Siderite occurs only outside of the contact zone. In the Wardner mines, which are several miles from the contact, siderite is most plentiful. The deposits were formed within the epoch of granitic intrusions, as shown by the occasional intersection of ore-bodies by lamprophyric dikes; and since the time of ore formation erosion has probably removed several thousand feet of rock.

The facts briefly set forth are of highest importance and serve to connect the high-temperature deposits with those of intermediate conditions.

O. H. Hershey believes that the metals of the Wardner district, in the western part of the Coeur d' Alene region, were originally disseminated in the Belt sediments and that these disseminations were eventually concentrated into ore-bodies by hot waters ascending on thrust faults. He also holds that the contact metamorphic deposit of the Success Mine has been formed prior to the monzonite intrusion and was invaded and metamorphosed by it.[1]

Lead-Silver Veins with Calcite, Siderite, and Barite.—Veins containing galena and zinc blende with a gangue of calcite, siderite, or barite are abundant in many mining regions and are frequently connected with replacement deposits in limestone. In many places they have a distinct connection with intrusive rocks and were formed shortly after the irruption, but some of them are similar to the Mississippi Valley lead-zinc deposits and may well have been deposited by the ascending waters of the ordinary circulation. Among numerous examples the deposits of Clausthal and Przibram may be briefly mentioned.

Lead-Silver Veins of Clausthal.[2]—The mines of Clausthal, in the Harz Mountains of Germany, which have been in operation since the thirteenth century and still maintain a moderate production, are working on a vein system which intersects a folded complex of Devonian and Carboniferous sedimentary beds, the prevailing rocks being clay, slate and graywacke. The general strike of the veins is east-west, and the dip is steep. The numerous veins extend over an area 15 miles in length and 5 miles in width; they are in general composite veins or lodes and the important fissures are also faults of considerable throw. Mining operations in this district have attained a depth of 3,000 feet.

The ores contain chiefly galena and zinc blende, with some marcasite, pyrite, chalcopyrite, and tetrahedrite. Arsenical minerals are generally absent. In one group of veins calcite and quartz predominate; in another barite and siderite. The galena contains only 0.05 per cent. silver, though richer ores with as much as 0.3 per cent. silver are found in some mines. Symmetrical banding is exceptional, the normal ore having an irregularly massive structure. The tendency appears to be toward an

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[1] J. B. Umpleby, Genesis of the Success zinc-lead deposits, *Econ. Geol.* vol. 12, 1917, pp. 138–153.

O. Hershey, *Idem*, pp. 348–558.

[2] A. von Groddeck, Ueber die Erzgänge des Oberharzes, *Zeitschr.* Deutsch geol. Gesell., 1866, pp. 693–776.

F. Klockmann, Beiträge zur Erzlagerstättenkunde des Oberharzes, *Zeitschr. prakt. Geol.*, 1893, pp. 466–471.

B. Baumgürtel, Oberharzer Gangbilder, Leipzig, 1907, pp. 23.

See also Stelzner and Bergeat, Die Erzlagerstätten, 1, 1905, pp. 763–771, and R. Beck, Lehre von den Erzlagerstätten, 1, 1909, p. 363–367.

increase in the percentage of zinc in depth. Within the lodes the clay slates are slightly altered by mechanical and chemical processes. The change in composition is slight[1] and appears to consist largely of an increase in sericite at the expense of an original chloritic mineral in the clay slate.

Opinions differ as to the genetic interpretation of the veins of Clausthal. There are no intrusive rocks in the immediate vicinity aside from a dike of kersantite, which is faulted by the vein fissures, and the mass of "Brocken" granite in the eastern part of the district. A genetic connection of these intrusives with the veins seems probable, but cannot be regarded as proved. According to von Koenen[2] the fissures are of Miocene age and some movement on these fissures seems to be taking place at present. The mineral association would indicate deposition at less depth or at lower temperature than in the veins of the Coeur d'Alene district, for instance.

Lead-Silver Veins of Przibram, Bohemia.[3]—The mines of Przibram, which have been worked for several hundred years and still maintain a small output, are situated 40 miles south-southwest of Prague, in the "Silurian syncline," well known in the early history of geology. The predominating rocks are Cambrian (?) graywacke and clay slate, a folded and faulted complex intruded by a stock of diorite. Dikes of diabase are exceedingly numerous and are followed by the veins; dikes of diorite and kersantite are also present. The intrusive diorite produces a decided contact metamorphism in Paleozoic sediments.

The veins have a steep dip and have been followed down to a depth of 3,773 feet; about forty of these veins have been worked, and they are contained within a narrow area 4 or 5 miles in length. The width of the veins attains 25 feet, but averages much less. Fig. 208 gives an idea of their structure. The ore minerals consist of galena and zinc blende with some pyrite and chalcopyrite and occasionally many other minerals like arsenopyrite,

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[1] A. v. Groddeck, *Jahrb.* Preuss. geol. Landes-Anstalt, 1885, pp. 1–52.
W. Lindgren, *Trans.* Am. Inst. Min. Eng., vol. 30, 1901, p. 683.

[2] A. von Koenen, Die Dislokationen W. und S. W. vom Harz., etc., *Jahrb.* Preuss. geol. Landes-Anstalt, 1903, pp. 68–82.

[3] J. Schmidt, Bilder von den Erzlagerstätten von Przibram. Published by Austrian Agricult. Dept., Vienna, 1887.
F. Posepny, *Archiv für prakt. Geol.*, 2, Freiberg, 1895, pp. 609–745.
A. Hofman and F. Slavik, Ueber Dürrerze von Przibram, *Bull. internat.* 15, Acad. Sci. de Bohème, 1910.

stibnite, uraninite, cobalt and nickel minerals, wurtzite, and millerite. Rich silver minerals like argentite and pyrargyrite, as well as native silver, were plentiful in the oxidized zone. The galena is the carrier of silver and contains about 0.5 per cent. of this metal. Among gangue minerals calcite, siderite, and quartz predominate, but barite and ankerite are also known. The structure is in part banded and drusy.

The quartz and zinc blende appear to increase in depth and the ores become "dry." These dry ores contain about 50 per cent. quartz, 17 per cent. siderite, 17 per cent. galena, 0.26 per cent. silver, and also primary boulangerite, tetrahedrite, pyrargyrite, diaphorite, specularite, chlorite, and cassiterite.

FIG. 208.—Section of the Adalbert vein at Przibram, Bohemia. *G*, Graywacke; *D*, diorite; *q*, quartz; *c*, calcite; *g*, galena; *b*, zinc blende. *After J. Zadrazil and J. Schmidt.*

The deep workings are practically dry, but there existed formerly a rich zone of oxidation descending, in spite of a high present water level, to depths of 200 to 900 feet.

The genetic connection of the veins with the intrusive diorite and its satellites of diabasic and lamprophyric dikes appears to be clearly indicated; the presence of cassiterite points in the same direction. That the region is a metallogenetic province connected with intrusions is also suggested by the occurrence of gold-bearing quartz veins in the diorite or granite.

Pyritic Galena-Quartz Veins.—In the surroundings of granitic and dioritic intrusions a certain type of lead-bearing veins is especially common, distinguished by pyrite, galena, dark zinc blende, arsenopyrite, and some chalcopyrite, with subordinate arsenopyrite, in a gangue of quartz, with a small amount of calcite or dolomite.

Freiberg, Saxony.—The type just mentioned corresponds closely to the "Kiesige Bleiformation" of Freiberg,[1] there represented by numerous veins of considerable persistency contained in a flat dome of biotite gneiss. K. Dalmer and other geologists have pointed out their probable genetic connection with the intrusive (Carboniferous) granites of the Erzgebirge and their well-established relationship to the tin veins which are situated closer to or within the intrusives. The Freiberg veins of this type are narrow, being seldom 3 feet wide, and have been mined to a depth of 2,100 feet. The pyrite, arsenopyrite, and zinc blende are poor in silver, but the galena contains 0.1 to 0.2 per cent. of this metal.

Stelzner and Schertel ascertained that the zinc blende contains microliths of cassiterite. The vein structure is irregularly massive, without marked banding or crustification.

The ores are of low grade, and, after a period of activity extending over nearly 750 years, the mines are now practically closed.

The silver-lead deposits of Freiberg comprise a complicated system of fissure veins of different types and ages, which have been carefully studied by such men as A. G. Werner (1791), A. von Weissenbach (1836), J. C. Freiesleben (1843), F. C. von Beust (1840), B. von Cotta (1861), and H. Müller (1849–1901).

The veins are classified as follows:

(1) Older Veins.—Noble[2] quartz formation: Fine-grained quartz with argentite, pyrargyrite, native silver, pyrite, and arsenopyrite.

Pyritic lead formation: Quartz, pyrite, galena, zinc blende, arsenopyrite, and chalcopyrite.

Tin formation: Quartz, fluorite, arsenopyrite, cassiterite, and chalcopyrite.

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[1] Herman Müller, Die Erzgänge des Freiberger Bergrevieres, *Erläuterungen zur geol. Special-Karte Sachsens*, Leipzig, 1901, p. 350.

[2] The word "Edel," or noble, refers to the high-grade silver ores.

Noble lead formation: Quartz, ankerite, rhodochrosite, galena, zinc blende, pyrite, tetrahedrite, pyrargyrite, proustite, and polybasite.

(2) Younger Veins.—Barytic lead formation: Barite, fluorite, quartz, calcite, galena (poor in silver), chalcopyrite, tetrahedrite, zinc blende. These veins are often of considerable width.

The barite veins are distinctly later than the older group, and their minerals occur in beautifully banded and drusy form. Müller is doubtless right in ascribing a Tertiary age to these veins and a possible connection with the basaltic eruptions of that age along the Bohemian frontier. The barytic lead veins sometimes carry nickel and cobalt minerals and Müller is inclined to correlate them with the cobalt and nickel veins of Annaberg.

The older group appears to be genetically connected with the granitic intrusions of Carboniferous age, or perhaps also with the Permian and Carboniferous porphyries (intrusive and effusive) of the same region. The "noble quartz formation" alone is intersected by dikes of quartz porphyry, while the other veins appear to be later than the porphyry. The granite stocks of the region are intersected by veins similar to those of Freiberg; but no granite occurs in the Freiberg district. It is interesting to note that in parts of the district dikes of kersantite and minette are plentiful and that the veins are later than these dikes.

Between the various members of the older group many transitions exist, and it seems justifiable to regard them as genetically connected with the granitic eruptions of Carboniferous age and as formed shortly after the last lamprophyric dikes of that parent magma had been intruded. The mineral association of the "noble quartz formation" and the "noble lead formation," with apparently primary argentite and pyrargyrite continuing to the greatest depth reached, far beyond the zone of oxidation, seems to suggest that these veins have been formed at relatively low temperature. They do not correspond to the types usually associated with intrusive masses.

The ore-shoots of the Freiberg veins are irregular; the richest parts were often at intersections of fissures (Fig. 72). The oxidized ores worked in the early history of the mines were rich in argentite and native silver.

Pyritic Galena-Quartz Veins in the United States.—Near intrusive areas in the central and eastern Cordilleran States are many veins of the Freiberg type, just described, although they ordinarily also carry gold together with silver. Few of them are, however, of the first importance. More common, perhaps, are veins which carry mainly massive galena and zinc blende associated with but little pyrite, or veins in which the pyrite entirely predominates. Examples of this kind are given in the description of the Leadville region (p. 613).

Excellent examples of the type, described by F. C. Schrader, occur near Kingman,[1] in northwestern Arizona, in the Wallapai mining district. These veins were discovered in 1872. Their upper parts yielded rich silver ores, but of late years the silver production has declined as the leaner primary sulphides were encountered, and in the ores now extracted zinc blende is the most valuable constituent. The greatest depth attained is about 1,400 feet. The rocks are pre-Cambrian granite, gneiss, and schist intruded by granite porphyry, probably of Mesozoic age, and by a great number of lamprophyric dikes of minette and vogesite, which in part are followed by the veins.

The deposits are well-defined fissure veins with steep dip, forming conjugated systems with northwesterly strike; they are straight and have well-defined walls, and some of them are traceable for considerable distances. The gangue is quartz, in places shattered and cemented by a later generation of calcite, occasionally also siderite. Among the primary sulphides are pyrite, galena, zinc blende, and chalcopyrite, rarely molybdenite and stibnite. The ore may contain $10 in gold and silver, 8 per cent. lead, and 5 to 16 per cent. zinc. It is in part shipped crude, in part concentrated.

The structure is irregularly massive, in places with rough banding by arrangement of the sulphides. The veins are narrow, though in some places ore-bodies 20 feet wide have been worked. The pay-shoots are irregular, but often coincide with intersections of veins. The water level is from 100 to 400 feet below the surface and above it were rich oxidized lead ores, horn silver, native silver, argentite, and ruby silver. The decrease of galena and increase of chalcopyrite noted in the lower levels suggest a gradual change in the primary filling.

The ore is mainly deposited by filling of cavities; the wall rocks contain little ore but are sericitized and filled with pyrite close to the veins.

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[1] F. C. Schrader, Mineral deposits of the Cerbat Range, Black Mountains, etc., *Bull.* 397, U. S. Geol. Survey, 1909.

THE SILVER-LEAD REPLACEMENT DEPOSITS IN LIMESTONE

General Features.—Limestone, dolomites, and calcareous shales are easily soluble by waters circulating above the water level along stratification planes, joints, veins, or zones of brecciation; caves and open passages will result. Below the water level more slowly circulating solutions often replace limestone by dolomite or cherty or jasperoid silica. If the solutions carry metallic sulphides these are easily precipitated, and by a simultaneous operation the carbonate goes into solution while a corresponding volume of sulphides takes its place (Fig. 29). Some of these replacement deposits that have no genetic connection with igneous rocks have been described above (p. 444).

In districts where metallization is caused by igneous activity the limestone is often replaced close to the contact by sulphides, particularly copper sulphides, associated with high-temperature minerals; these deposits are described in Chapter XXVI. Frequently, however, replacement by sulphides is also found at greater distances from the igneous rock, but the circulating solutions which caused the replacement, while probably derived from the magma, had a lower temperature and therefore no high-temperature minerals could form. Such deposits, which contain mainly lead, zinc, and silver, may appear in connection with eruptions of lavas and may form relatively close to the surface, but they are more common in the vicinity of intrusive rocks now exposed by erosion. The process is therefore favored by higher temperature and pressure.

For the development of replacement deposits, pathways that can be followed by the solutions are necessary. Joints and seams may provide them, but more commonly the fissures which were formed during or after the intrusion guide the solutions to the limestone. When the waters have entered a fissure the processes of replacement begin immediately but the products of interchange are not confined to this fracture. On the contrary, they spread in all directions, guided by minor structural planes, and replacement deposits in limestone are therefore characteristically irregular; it often happens that the original fissure may be difficult to discover, though genetically it is the key to the extent and the continuation of the deposit. The mining of such deposits demands thorough knowledge of the geological structure.

There are a great number of such deposits in districts of the

Cordilleran region of the Americas. Many of them are small and are soon exhausted, while others are among the great ore deposits of the world. The districts of Aspen and Leadville, Colorado; Eureka, Nevada; Lake Valley, New Mexico; Elkhorn, Montana; Park City and Tintic, Utah; and Sierra Mojada, Mexico, may serve as examples.

At some places these silver-lead deposits follow dikes or intrusive sheets, but such deposits were usually formed after the rock had congealed and cooled. At other places they are dependent upon impervious overlying beds like shale. The latter mode of occurrence is exceedingly common (Fig. 70) and indicates that the solutions were ascending and that deposition followed the ponding or stagnation of the solution or at least was favored by less rapid circulation. Sections of two smaller replacement deposits are shown in Figs. 209 and 210.

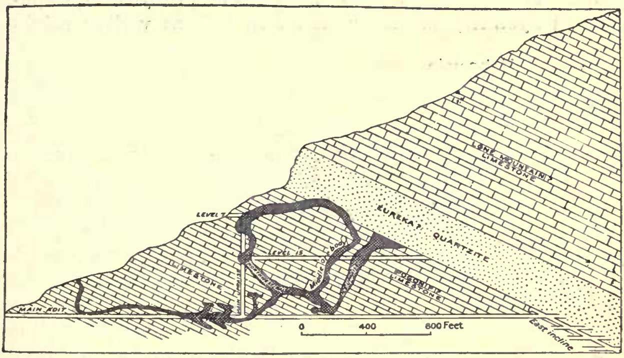

FIG. 209.—Irregular replacement deposit in the Garrison mine, Cortez, Nevada. Ore consists of galena, zinc blende, pyrite, stromeyerite, etc., and their oxidation products. *After W. H. Emmons, U. S. Geol. Survey.*

The primary minerals of these replacement deposits are comparatively few and simple. Deep oxidation is, however, common in limestone and descending waters may effect many changes and develop a great number of rare oxidized minerals in the oxidized zone, while complex secondary sulphides may form in the lower parts of the deposit. The gangue minerals are few: Dolomite is often present as a coarser aggregate and at many places the process of replacement was begun by a dolomitization of the limestone. Dense, cherty quartz is exceedingly common, much more so than coarser crystalline quartz. In accordance with the suggestion of Spurr, this siliceous gangue is called jasperoid, though this term is really a misnomer, for the rocks are gray rather than red or brown. Other gangue minerals are calcite, barite, sometimes fluorite, various carbonates allied to ankerite, and more rarely rhodochrosite. The replacement deposits carrying siderite are described on page 595. The common primary ore minerals are pyrite, galena, zinc blende, chalcopyrite, and more rarely arsenopyrite. Tetrahedrite, tennantite, enargite, bornite, bismuthinite, wolframite, molybdenite, and stibnite are of local importance. But when argentite, ruby silver, stephanite, polybasite, and native silver as well as various sulphantimonides of lead appear the probability is that they are secondary minerals. Chalcocite is in these deposits probably always secondary. Gold is sometimes present as a primary mineral, but the ores carry ordinarily much more silver than gold. Galena is very common and is usually rich in silver. The silver content of galena is usually caused by a primary intergrowth with small grains of argentite. In many so-called lead deposits the lead really predominates only in the oxidized zone, while the primary ore carries far more pyrite and zinc blende than galena. Such are the relations at Leadville, for instance.

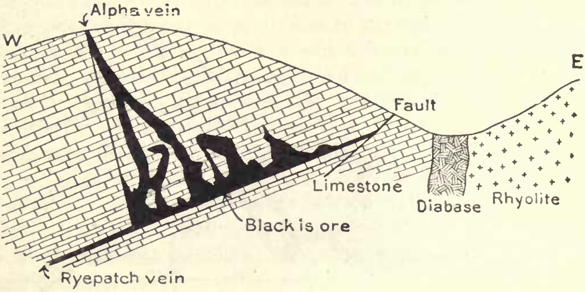

FIG. 210.—Section of Ryepatch mine, Unionville, Nevada. Ore-body between faults 250 feet wide; consists of calcite, quartz, pyrite, galena, zinc blende, tetrahedrite, etc. *After F. L. Ransome, U. S. Geol. Survey.*

The ore which replaces limestone is usually coarse-grained, while, as mentioned above, the replacement of limestone by silica yields rocks with fine grain. Crustified or drusy structures are unusual, though common in the oxidized parts of these deposits. Before the importance of replacement as a geological process was recognized many of these ores were considered as fillings of limestone caves. Some of the deposits consist of massive sulphides, while in others, presumably formed at lower temperature, the gangue may prevail.

Replacement deposits are not confined to calcareous rocks. They occur also in quartzite, shale, and igneous rocks, but they are certainly more common in carbonate rocks than elsewhere. Very hot solutions may replace any rock, but most of the deposits described in this chapter were probably laid down by solutions having a temperature of less than 200° C., and under such circumstances limestone would be replaced while other rocks would be little affected. Siliceous rocks are more easily replaced than aluminous material; it is evidently difficult to carry away large amounts of alumina even at high temperatures, and while in limestone the replacement is often complete, ores in aluminous rocks contain much residual material.

Park City, Utah.[1]—The Park City district lies near the summit of the Wasatch Range. Since 1870 it has yielded silver to the value of $97,000,000 and lead valued at $53,500,000, lately also much copper and zinc and it still remains one of the most important metal-producing districts of the United States. Its ores are in part shipped as mined, but much is also concentrated, the total output of crude ore for 1917 being 240,600 tons. The concentrating ore contains from 6 to 8 per cent. lead, 6 to 8 per cent. zinc, 6 to 10 per cent. iron, and 9 ounces of silver per ton; also some gold and copper. The deepest shafts attain 1,500 and 2,000 feet and the workings of the district probably aggregate 100 miles in length.

A huge anticline of late Carboniferous, Permian, and Triassic sediments, mainly limestone, quartzite, and shale, the total thickness of beds exceeding 8,000 feet, is intruded by laccolithic stocks of diorite porphyry, probably of late Cretaceous age, which have caused contact metamorphism in the adjoining limestone and shales.

The ores occur as lode deposits and closely associated bedded

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[1] J. M. Boutwell, *Prof. Paper* 77, U. S. Geol. Survey, 1912.

V. C. Heikes, Mine production of Utah, in *Mineral Resources*, U. S. Geol. Survey, pt. 1, Annual publication.

deposits, in two parallel zones extending northeastward. The bedded deposits, mainly in limestone, have been mined to a depth of 900 feet; the lode deposits continue to the greatest depths attained. The lode deposits intersect the sediments and the porphyry as well, have a steep dip, and often lie in quartzite or between limestone and quartzite. The ores are in part deposited by filling of seams in shattered ground, in part by replacement. The stopes are as much as 30 feet in width.

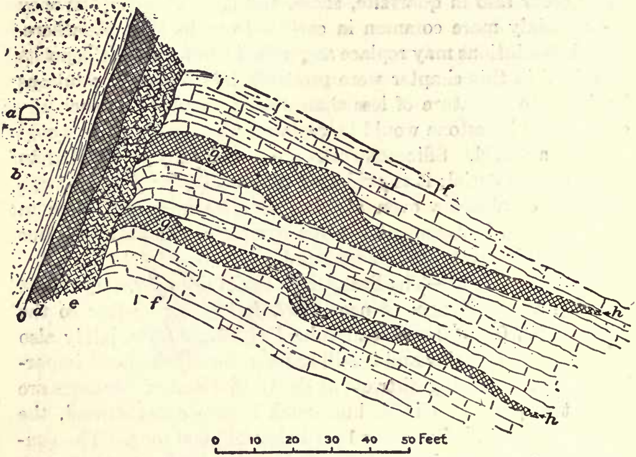

FIG. 211.—Vertical section of rich lead ore occurring in veins and in replacement deposits, Kearns-Keith mine, Park City, Utah. *a*, Tunnel; *b*, diorite porphyry, sheeted and pyritic; *c*, hanging-wall fissure; *d*, lead ore in siliceous gangue; *e*, breccia zone with ore fragments; *f*, marmorized limestone, Thaynes formation; *g*, *h*, banded replacement ore, in part oxidized. *After J. M. Boutwell, U. S. Geol. Survey.*

The bedded deposits are massive sulphides replacing limestone strata in two of the calcareous formations and are from a few inches to 10 feet thick, 500 to 800 feet in the direction of the strike and at most 200 feet along the dip. The relation between the two types is shown in Fig. 211. The layers of the bedded ore are made up of ore and gangue minerals in granular texture exactly like that of the original limestone. There is evidence of two epochs of deposition, for some of the bedded ores near the porphyry contacts contain garnet with calcite as gangue, while the lode deposits and the bedded ores associated with them are free from garnet and were formed after the cooling of the porphyry.

The ore minerals are galena, zinc blende, tetrahedrite, and a little chalcopyrite. Tetrahedrite is often intergrown with coarse galena. The gangue is mainly quartz and jasperoid; fluorite, calcite, and rhodonite occur locally. Sericitization is noted where the lodes intersect porphyry. The richest ore was formed in the bedded deposits; the ore in depth is of leaner grade, but carries more copper and zinc.

The Park City mines are very wet and the water level is high. In view of this it is remarkable that the oxidation is deep and partial oxidation has been noted to a depth of 1,200 feet. The oxidized zone contained apparently but little native silver and cerargyrite.

Tintic, Utah.[1]—The replacement deposits of the Tintic district, situated in a desert range 70 miles south of Salt Lake City, exemplify another type, which has been so modified by oxidation that the original character of the ore is sometimes difficult to interpret. Paleozoic limestones are intruded by a monzonite stock that formed the core of a volcano of early Tertiary age, the surface flows of which are largely eroded. A number of narrow fissures traverse both monzonite and limestone; in the former the deposits are pyritic veins with sericitized walls, while in the limestone the inconspicuous fractures widen locally into large or small ore bodies characterized as chambers, chimneys, pipes, pockets or pods, which in large part replace the adjacent rock or follow, for a distance, stratification planes, fissures or joints (Fig. 212). In the Iron Blossom mines the galena ore forms a horizontal pipe-like ore shoot with a greatest width of 150 feet. This has been mined for nearly 8,000 feet and follows

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[1] G. W. Tower, Jr., and G. O. Smith, Geology and mining industry of the Tintic district, Utah, *Nineteenth Ann. Rept.* U. S. Geol. Survey, pt. 3, 1898, pp. 603–785.

W. P. Jenney, The mineral crest, etc., *Trans.* Am. Inst. Min. Eng., vol. 33, 1903, pp. 46–50; 475–483.

V. C. Heikes, *Op. cit.*

W. Lindgren and G. F. Loughlin, *Prof. Paper* 107, U. S. Geol. Survey, 1919.

W. Lindgren, Processes of mineralization and enrichment in the Tintic district, *Econ. Geol.*, vol. 10, 1915, pp. 225–240.

G. W. Crane, Geology of the ore deposits of the Tintic district, *Trans.*, Am. Inst. Min. Eng., vol. 54, 1917, pp. 342–355.

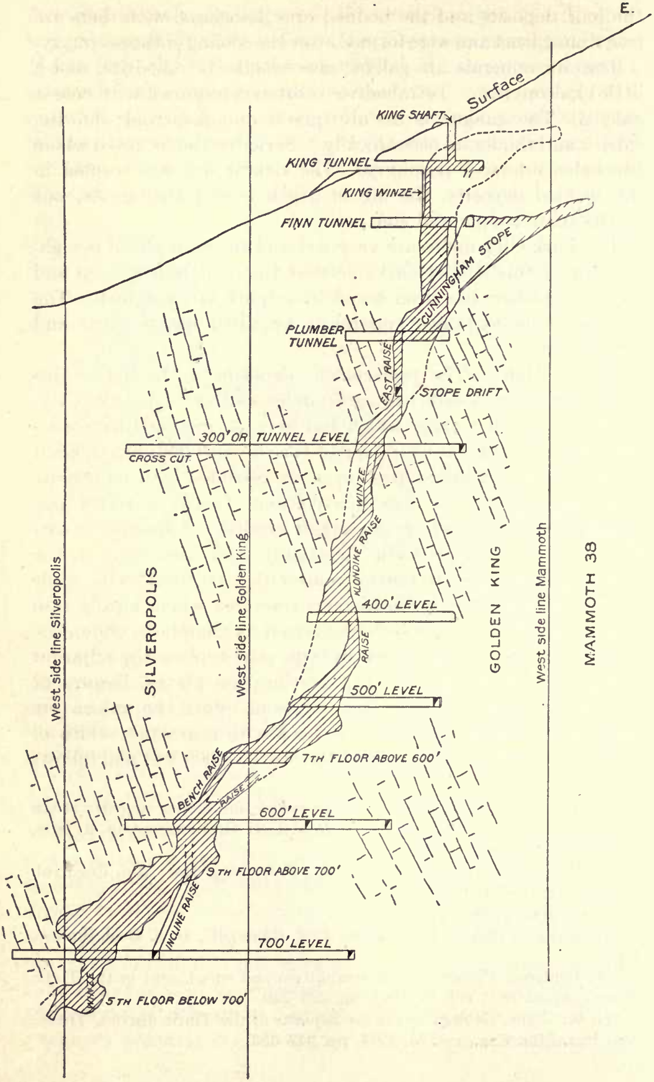

FIG. 212.—Vertical section showing Silveropolis ore column, Mammoth mine, Tintic, Utah.

the intersection of an obscure fissure with a certain bed of pure limestone.

The primary ore minerals are galena and zinc blende with very little pyrite. The gangue minerals are fine grained quartz or jasperoid and barite. The galena is rich in silver (20–50 ounces per ton). Other large replacement bodies consist mainly of quartz and jasperoid and carry gold and silver with a little lead. Near the intrusive mass the ore contains mainly enargite (Cu_3AsS_4) with a little gold.

The jasperoids along the deposits result from the replacement of limestone and dolomite by colloidal silica, which later crystallized to chalcedony and quartz. The crushing of this silica resulted in brecciation and silica of a second phase was deposited in open spaces as colloid material which later crystallized to quartz. The development of crystalline barite, galena and other sulphides accompanied both phases.

Complete or partial oxidation continues to a depth of about 2,000 feet, which coincides with the water level. In the mines in monzonite the water level stands much higher.

During the deep oxidation cerussite and anglesite formed in the lead deposits. Secondary zinc minerals, mainly smithsonite, usually develop in the limestone outside of the primary lead shoots.[1] Complex copper and copper-lime arsenates in the copper bearing deposits and these minerals are usually accompanied by more or less chalcocite and covellite.

The mines of Tintic yield annualy $6,000,000 to $9,000,000, their complex smelting ores containing gold, silver, copper, lead, and zinc. The annual ore production is about 300,000 tons. The total value of the silver, lead, gold and copper produced from 1869 to 1916, inclusive, is approximately $170,000,000.

Aspen, Colorado.[2]—The ore deposits at Aspen, in the central part of Colorado (Fig. 173), for many years yielded a large amount of lead and silver, and the annual output is still of considerable value. During recent years zinc blende has been added to the products of this district. The ores average 5 ounces of

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[1] G. F. Loughlin, The oxidized zinc ores of the Tintic district, *Econ. Geol.*, vol. 9, 1914, pp. 1–19.

[2] J. E. Spurr, *Mon.* 31, U. S. Geol. Survey, 1898.

J. E. Spurr, Ore deposition at Aspen, Colo., *Econ. Geol.*, vol. 4, 1909, pp. 301–320.

silver per ton and 6¼ per cent. lead. The geological column at Aspen includes 200 to 400 feet of Cambrian quartzite; 250 to 400 feet of Silurian dolomite; 60 feet of Devonian quartzite and shale ("Parting quartzite"); 250 feet of lower Carboniferous dolomite and 150 feet of limestone of the same age; 1,000 feet or more of thin-bedded Carboniferous limestones and shales called the Weber formation; and a great thickness of Carboniferous, Triassic, and Cretaceous sandy and calcareous sediments. The entire series is sharply upturned. A sheet of diorite porphyry intrudes the lower Paleozoic formations and a sheet of rhyolite porphyry lies at the base of the Weber formation. Both intrusives are of late Cretaceous or early Tertiary age. Complicated faulting and local doming accompanied the intrusion. During the short epoch of ore deposition sulphides were deposited along the faults and fractures, the most important horizon of mineralization being at the base of the Weber shales, where the depositing waters were dammed by the relatively impervious shale and it may also have acted as a precipitant. In his later paper Spurr differentiates the complicated deposits as (1) barite veins; (2) silver sulphides, sulphantimonides, and sulpharsenides; (3) galena and zinc blende veins. This series is believed to have been deposited under conditions of gradually rising temperature. Faulting continued after the short epoch of ore deposition.

The barite veins are generally barren. After their development rich sulphides such as argentite, polybasite, and tetrahedrite were deposited. A remarkable shoot of polybasite ore yielding many million dollars was mined in the Molly Gibson mine, at a depth of a few hundred feet below the surface. This occurrence of rich silver sulphides certainly suggests enrichment by descending surface waters, but Spurr insists on its primary origin. The last phase of mineralization consisted in the deposition along zones of fracture and brecciation of lead and zinc ores of milling grade and poor in silver. The ore minerals occur in disseminated form, replacing limestone, either without gangue or with a little jasperoid and dolomite.

The mine water is abundant and the water level stood originally about 300 feet below the surface. Down to 1,000 feet (the greatest depth thus far reached) native silver with some barite has been deposited by descending solutions, probably by the reducing influence of the carbonaceous Weber shales.

Leadville, Colorado.[1]—Since the discovery of the ore deposits of Leadville, Colorado, in 1875 this district has yielded an enormous production of lead and silver, also much gold, copper, and zinc. Previous to that date placers were worked in the district and gold to the value of several million dollars was washed from the gravel of the gulches. For a long time after 1875 the oxidized lead ores, containing much iron and manganese, were worked. At the present time the main product is heavy sulphide ore containing pyrite, zinc blende, galena, and chalcopyrite. Bodies

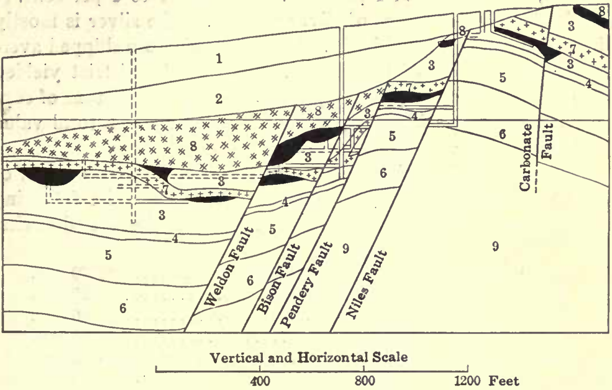

FIG. 213.—Vertical section showing geological structure and occurrence of ore-bodies at Leadville, Colorado. 1, Wash; 2, lake beds; 3, Leadville blue limestone (Carboniferous); 4, parting quartzite (Devonian); 5, white limestone (Silurian); 6, lower quartzite (Cambrian); 7, gray porphyry (early Tertiary); 8, white porphyry (early Tertiary); 9, granite (pre-Cambrian); ore-bodies in black. *After S. F. Emmons and J. D. Irving, U. S. Geol. Survey.*

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[1] S. F. Emmons, Geology and mining industry of Leadville, Colorado, *Mon.* 12, U. S. Geol. Survey, 1886.

S. F. Emmons and J. D. Irving, The Downtown district, *Bull.* 320, U. S. Geol. Survey, 1907.

A. A. Blow, *Trans.* Am. Inst. Min. Eng., vol. 18, 1890, pp. 145–181.

Philip Argall, *Eng. and Min. Jour.*, vol. 89, 1910, p. 261.

Philip Argall, *Min. and Sci. Press*, July 11 and 15, 1914.

Max Boehmer, *Trans.* Am. Inst. Min. Eng., vol. 41, 1911, pp. 162–165.

C. J. Moore, *Econ. Geol.*, vol. 7, 1912, pp. 590–592.

J. D. Irving, *Prof. Paper*, in course of publication, U. S. Geol. Survey.

of oxidized zinc minerals such as calamine and smithsonite have recently been discovered.[1]

Out of a total of 477,000 tons mined in 1916, 310,000 tons were sulphide ores, and 18,000 tons iron-manganese ores of the oxidized type containing a little silver and used for flux. The heavy sulphide ores are in part smelted directly and in part concentrated; they vary considerably but consist mainly of pyrite and zinc blende, with less than 1 per cent. of copper, 1 to 4 per cent. of lead, and 2 to 9 ounces of silver per ton. The silver is mostly in the galena and zinc blende. The crude zinc ore shipped averaged 30 per cent. of that metal. In 1916 the district yielded $1,470,000 in gold, 2,784,000 ounces of silver, 1,300 tons of copper, 11,000 tons of lead, and 38,000 tons of zinc—a total value of $15,600,000.[2]

The geological section consists, according to S. F. Emmons, of Paleozoic rocks resting on granite and gneiss. The following formations are important in the study of the ore deposits.

Weber shales and grits, lower Carboniferous	2,500 feet.
Blue limestone, lower Carboniferous	200 feet.
Parting quartzite, Devonian	40 feet.
White limestone, Silurian	160 feet.
Lower quartzite, Cambrian	150 to 200 feet.

These formations are intruded by numerous sheets of porphyry, which in the main lie parallel to the bedding, but in places cut diagonally across it. Some sheets are thin, and others are nearly 1,000 feet in thickness. The "white porphyry" is a siliceous granite porphyry, which Spurr[3] calls alaskite porphyry. The "gray porphyry" is similar but contains remains of resorbed ferromagnesium minerals and is a little lower in silica. The white porphyry is normally intruded in the blue or Leadville limestone; the gray porphyry forms thinner sheets at various horizons.

The intrusions and the ore deposition were followed by a marked doming of the strata and faulting of great complexity, so that the district now consists of numerous blocks successively dropping off toward the Arkansas Valley (Fig. 213). The

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[1] G. M. Butler, Some recent developments at Leadville, *Econ. Geol.*, vol. 7, 1912, pp. 315–323; vol. 8, 1913, pp. 1–18.

[2] C. W. Henderson, Mines report of Colorado in *Mineral Resources*, U. S. Geol. Survey, pt. 1, Annual publication.

[3] *Prof. Paper* 63, U. S. Geol. Survey, 1908, p. 70.

ore deposits are found mainly in the blue limestone at or near the contact with the overlying white porphyry (Fig. 214). The upper surface of the ore is often remarkably regular and sharp, being formed by the porphyry contact, while the lower surface is irregular. Although this is the normal development, replacement ores are also found in other positions, along fissures in limestone (Fig. 215), along fault planes or below the gray porphyry or in fissure veins extending below the sedimentary beds. The fissure veins are confined to the part of the district near the Ibex mine; many of them are rich in native gold.

Some of the ore-bodies are of large size, especially in a horizontal direction. Owing to their mode of occurrence and to the great quantity of mine water a depth of only 1,500 feet has been attained; it was considered useless to go below the basement of the Paleozoic formations.

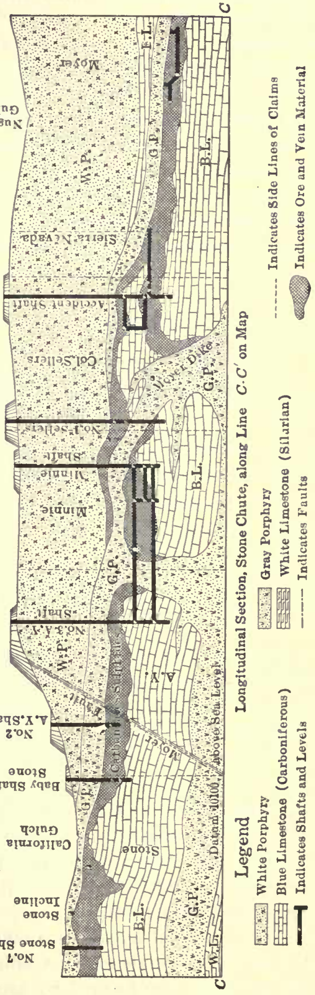

FIG. 214.—Longitudinal section of Stone ore-shoot, Iron Hill, Leadville, Colorado. *After A. A. Blow.*

Blow has shown that the ore-shoots on Iron Hill follow northeastward-trending zones parallel with cross cutting sheets of gray porphyry.

The usual ore is a massive granular mixture of sulphides, among which pyrite and zinc blende prevail. There is a scant gangue of quartz, jasperoid, and little barite. The limestone near the deposits often contain much manganosiderite spreading from the ore-bodies. The contact between ore and limestone is usually surprisingly sharp, though irregular. Among the rarer minerals is native gold, molybdenite, wolframite and scheelite; the ores contain a little antimony, arsenic, and bismuth, and the presence of traces of tellurium in the pyrite has been shown by Pearce.

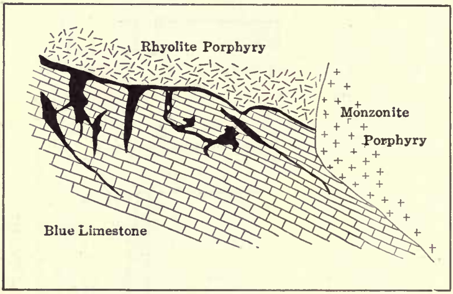

FIG. 215.—Cross-section of shoot at Oro La Plata mine, Leadville, Colorado, showing irregular ore-bodies along fractures. *After J. D. Irving.*

The oxidized ores of the upper levels which are still mined to some extent contain limonite, manganese oxide, cerussite, and zinc carbonates. They were frequently rich in silver chloride.

The genesis of the Leadville ores has been discussed extensively. Emmons held that they were formed by aqueous solutions coming from above and that they derived their mineral content mainly from the igneous rocks, but he did not deny the possibility that the solutions may originally have come from great depths, nor did he assert that they were necessarily derived from the eruptive rocks in immediate contact with the deposits. He also fully recognized that at the time of ore formation the present deposits were covered by about 10,000 feet of overlying rocks. Other writers, among them A. A. Blow, have sought to prove that the deposits were formed by ascending solutions.

The deposits of Leadville are unusual in that the sulphide replacement is so complete and that the contacts with the limestone and porphyry are so sharp. They strongly resemble the contact-metamorphic deposits except in the association of gangue minerals, which points clearly to moderate temperatures at which calcium silicates could not form, and this seems to prove that the ore deposition did not take place immediately after the intrusion. The recent discoveries of vein deposits near the Ibex Mine and ores containing magnetite and epidote also point to deep-seated sources for the metals far below the present ore horizon; but how the solution could penetrate along the contacts for so long a distance without visible passageways is as yet a mystery.

The Leadville-Boulder County Belt.—The Leadville deposits form only a single unit in a belt of deposits which extends for 80 miles in a northeasterly direction and comprises a great many districts, including the Kokomo, Alma, Fairplay, Breckenridge, Montezuma, and Argentine and continuing through Clear Creek, Gilpin, and Boulder counties (Fig. 173). The deposits include replacement bodies and veins and are found in rocks of the most diverse kinds. A common feature of the whole belt is a series of intrusives, appearing as sheets in the sedimentary formations and dikes or smaller stocks in the pre-Cambrian granite and schists. The inference that these intrusives are genetically connected with the deposits seems well founded. S. F. Emmons and Whitman Cross first called attention to this belt of intrusives; S. H. Ball[1] and F. L. Ransome[2] have discussed the petrography of the porphyries and Ransome has presented a diagram showing the composition of all analyzed varieties. The most abundant intrusives are alaskite porphyry, granite porphyry, bostonite porphyry, monzonite porphyry, and quartz monzonite porphyry. The dikes commonly extend in a northwesterly direction, but show no great individual continuity. Ball has indicated on a map[3] all the occurrences of porphyry within this belt. Emmons held that the intrusions were probably of Jurassic age, but later evidence discovered by Cross and others has shown that they must be later, falling either in the very latest Cretaceous or the earliest Tertiary. The ore deposits are later than the porphyries but were probably formed shortly after their consolidation.

[1] *Prof. Paper* 63, U. S. Geol. Survey, 1908, pp. 67–70.

[2] *Prof. Paper* 75, U. S. Geol. Survey, 1911, pp. 60–62.

[3] *Op. cit.*, Pl. XI.

The prevailing type is a sulphide ore with abundant pyrite and zinc blende and lesser amounts of galena and tetrahedrite. Chalcopyrite is subordinate, arsenopyrite rare. Telluride ores occur occasionally in the eastern end of the belt. Silver prevails in the southeastern part and gold is the important metal in Gilpin and Boulder counties. The gangue is made up of quartz, siderite, manganosiderite and other carbonates, but not much rhodochrosite is present. In places there is considerable barite.

The replacement deposits of Leadville, in Carboniferous limestone below porphyry sheets, have already been mentioned. In the Tenmile district[1] replacement deposits and fissures veins appear in the upper Carboniferous formations, with pyrite, zinc blende, and galena in a gangue of quartz, calcite, rhodochrosite, and barite. The Red Cliff district[2] has ores similar to those of Leadville, also replacement deposits carrying gold in Cambrian quartzite. The Breckenridge district[3] contains fissure veins intersecting Cretaceous shale and monzonite sheets, with pyrite, zinc blende, and galena as the principal ore minerals. Remarkable pockets of crystallized gold are thought to be deposited by descending waters. Northeast of Breckenridge is the Montezuma district,[4] in pre-Cambrian rocks with northeasterly trending veins carrying pyrite, chalcopyrite, galena, and zinc blende, in places with ruby silver or similar rich silver minerals, which are apparently of later origin than the rest of the ore. The gangue consists of quartz, siderite, and barite. Farther up, at the Continental Divide, is the Argentine district, with veins in gneiss, which carry similar ore minerals in a gangue of quartz, calcite, and fluorite. These veins contain both silver and gold.

The Clear Creek district[5] lies also in the pre-Cambrian area. The rocks comprise an older division of sedimentary origin, the Idaho Springs formation, intruded by granite, diorite, and pegmatite. A complex system of dikes of the kinds mentioned above is followed by the veins, which are principally silver deposits containing galena, zinc blende, pyrite, tetrahedrite, and chalcopyrite, in a more or less scant gangue of earlier

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[1] S. F. Emmons, *Folio* 48, *U. S. Geologic Atlas*, 1898.

[2] A. H. Means, *Econ. Geol.*, vol. 10, 1915, pp. 1–27.

[3] F. L. Ransome, *Prof. Paper* 75, U. S. Geol. Survey, 1911.

[4] H. B. Patton, *First Rept.*, Colo. Geol. Survey, 1909, pp. 112–144.

[5] J. E. Spurr, G. H. Garrey, and S. H. Ball, Economic geology of the Georgetown quadrangle, *Prof. Paper* 63, U. S. Geol. Survey, 1908.

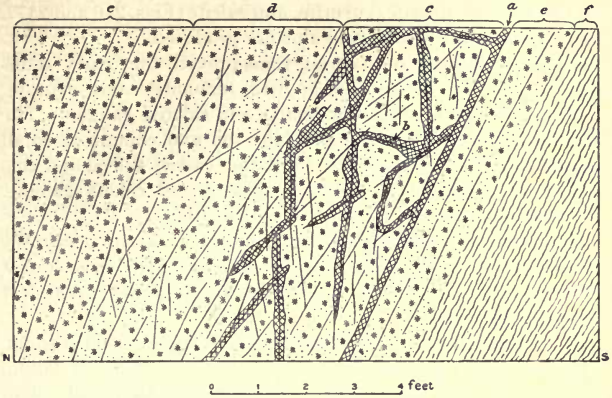

FIG. 216.—Section of Pelican vein, Georgetown, Colorado. *a*, Main vein; *b*, ore, mainly zinc blende; *c*, ore zone of brecciated and silicified alaskite porphyry; *d*, fractured and silicified porphyry; *e*, sheeted porphyry; *f*, altered gneiss. *After J. E. Spurr, U. S. Geol. Survey.*

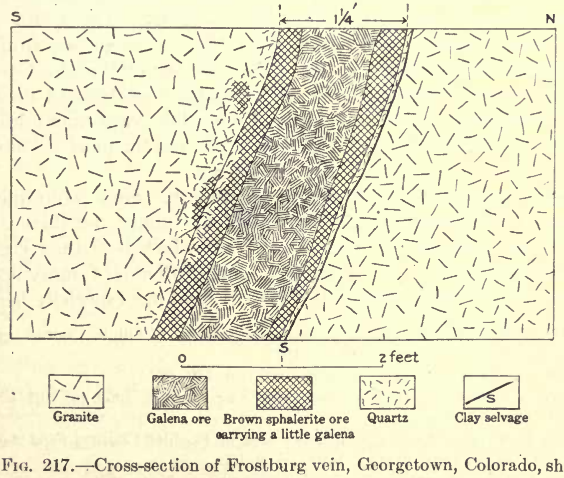

FIG. 217.—Cross-section of Frostburg vein, Georgetown, Colorado, showing deposition of galena and zinc blende. *After J. E. Spurr, U. S. Geol. Survey.*

quartz and later siderite, ankerite, and calcite (Figs. 216 and 217). Silver, gold, copper, lead, and zinc are produced.

In Gilpin County[1] gold-bearing veins prevail; they carry abundant pyrite, with some chalcopyrite, tennantite and enargite, rarely pitchblende, in a scant quartz and siderite gangue. The ores average about $8 in gold to the ton and contain a small amount of silver. A later vein formation carries galena and zinc blende, with silver, in a quartz, siderite and calcite gangue. Quartz-telluride ores are also known (Fig. 41).

In all the districts the alteration of the feldspathic country rock adjacent to the vein is of the sericitic type. The lowest workings in Clear Creek and Gilpin counties are 2,000 feet below the outcrops.

In Boulder County, the present production of which is small, there are some gold-bearing veins similar to those of Gilpin County, and also some veins which have produced rich silver ores, but the most interesting types are the telluride veins, which are rare in the other districts mentioned, and the tungsten veins, which are absent elsewhere.

The Enterprise vein,[2] which is typical of the telluride deposits, consists of several narrow seams, forming a sheeted zone along which filling and replacement have occurred. The width of this zone is from 1 to 3 feet and the filling is often beautifully banded with abundant druses. The country rock is pre-Cambrian granite. The minerals are crystallized quartz and dense jasperoids with barite, adularia, and roscoelite (vanadium mica); there is a little pyrite and much molybdenite; the most valuable minerals are the gold and silver tellurides.

The Tungsten Deposits of Boulder County.—Some wolframite occurs in the gold-bearing veins of Boulder County, but there is a fairly well-defined area near Nederland[3] which is characterized by tungsten ores. The tungsten mining in Boulder County is of recent origin but has developed rapidly, and has especially been

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[1] E. S. Bastin and J. M. Hill, Economic geology of Gilpin County, etc. *Prof. Paper* 94, U. S. Geol. Survey, 1917.

[2] W. Lindgren, *Econ. Geol.*, vol. 2, 1907, pp. 453–463.

T. A. Rickard, *Trans.* Am. Inst. Min. Eng., vol. 33, 1903, pp. 567–577.

[3] W. Lindgren, *Op. cit.*

R. D. George, The main tungsten area of Boulder County, *First Ann. Rept.*, Colorado Geol. Survey, 1908, pp. 9–103.

F. L. Hess, chapter on production of tungsten, *Mineral Resources*, U. S. Geol. Survey, Annual publication.

stimulated by war conditions. In 1918, 5,020 tons of tungsten concentrate were produced in the United States of which the larger part came from the deposits of Boulder County. The remainder of the domestic production comes largely from Atolia, Kern County, California, where scheelite ($CaWO_4$) occurs in veins associated with gold-bearing quartz veins of the middle depths. Queensland, Argentina, Portugal and Burma furnish the bulk of the foreign production. The ore from these four countries comes from high-temperature veins related to the cassiterite veins. For concentrates with 60 per cent. WO_3 the price in 1910 was from $400 to $500 per ton; in 1918 about $1,200 per ton. Tungsten is used mainly for high-speed tool steel, the alloy ferro-tungsten or metallic tungsten being first produced. It is also used in incandescent lamps, etc. (p. 673).

The veins of Boulder County are narrow, often brecciated fissures in granite, some of them following porphyry dikes. The gangue is made up of quartz and fine-grained silica. Kaolin is abundant as a secondary mineral. There is a little pyrite, but the principal mineral is the iron tungstate, ferberite, which occurs in a fine-grained mixture with quartz or as beautiful crystals coating druses and also associated with quartz crystals.

Summary.—The mineral belt described in the preceding paragraphs illustrates well the intimate connection between veins and replacement deposits. With all their differences the deposits are evidently of approximately contemporaneous origin. The southwest end at Leadville represents the deposits formed at considerable depth, probably not less than 10,000 feet as estimated by Emmons. Toward the northeast end the sedimentary series is absent and probably never covered the Front Range of Colorado. There is evidence in this range of a tertiary surface higher than the present but of less relief; and Spurr believes that effusive rocks once rested on this surface. S. H. Ball[1] estimates that in Clear Creek County the total erosion since the intrusion of the porphyries has been about 5,000 feet, and it is probable that it was less in Boulder County. Corresponding to this difference in geologic conditions is a difference in structure and composition of the ore deposits. At the southwestern end of the belt heavy sulphide ores prevail, mostly as replacements. In Clear Creek and Gilpin counties the banded and drusy structure of the veins begins to be apparent, and in Boulder County we find

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[1] *Prof. Paper* 63, U. S. Geol. Survey, 1908, p. 145.

veins like the telluride deposits and the tungsten veins, which distinctly resemble the deposits formed at slight depth below the surface, having drusy, banded structure and containing fine-grained quartz and tellurides. These relations are, to say the least, very suggestive of progressive change in original depth of deposition from the western to the eastern end of the belt.

The rich silver minerals, which are found in some of the districts, are regarded by Spurr and also by E. S. Bastin[1] as products of deposition by descending surface waters. Spurr believes that the metals in the Clear Creek deposits were derived from deep-seated magmatic sources and, dissolved in magmatic waters, ascended the fissures after each intrusion of porphyry. He also believes that most of the gangue minerals—quartz, carbonates, and barite—were formed through the leaching of the country rock by the ascending waters.

Ransome is less positive regarding the genesis of the deposit at Breckenridge and points out that many of the fissures are short along the strike and die out before reaching great depth, so that instead of directly ascending solutions there was probably considerable lateral movement during which the solutions had opportunity to search the surrounding rocks more or less thoroughly. The gangue minerals are also regarded by him as derived from the country rock, while he considers a magmatic origin possible for the metals and elements like sulphur and fluorine. In view of the faint contact-metamorphic effects shown at the western end of the belt in the limestones, it is probable that little if any of the mineral-bearing solution was derived from the small bodies of intruding porphyries and that the metals were rather derived from deep-seated magma basins.

DEPOSITS WITH NATIVE SILVER

Native silver is not, as a rule, a primary mineral in the deposits which contain it, nor is it restricted to any particular class of deposits. As a secondary product due to reactions within the oxidized zone it is common in many kinds of deposits—for instance, in argentiferous galena ores, in tetrahedrite ores, and in the argentite veins in the Tertiary lavas. It is ordinarily found some distance below the surface; cerargyrite (AgCl) is more abundant in the outcrops. The native silver often occurs at

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[1] *Econ. Geol.*, vol. 8, 1913, p. 51.

depths far below the zone of oxidation, properly speaking. At Aspen, Colorado, it is abundant in fissures and vugs of limestone and shale 900 feet below the surface and is distinctly later than the primary lean galena-zinc blende ores; along the delicate threads of the metal small barite crystals are often suspended.

In certain deposits the native silver is the predominating ore mineral down to considerable depths. Some of these occurrences are described below and may be divided into two groups:

1. The silver-bearing deposits with zeolitic enrichment.
2. The silver-bearing cobalt-nickel deposits.

The Zeolitic Enrichments.—Zeolites are ordinarily foreign to ore deposits connected with igneous rocks. Their occurrence in some contact-metamorphic deposits and in fissure veins is mentioned on page 427. Zeolites are found in the deposits of Kongsberg, Norway; Andreasberg, Germany; the Arqueros and other mines in Chile; and Guanajuato, Mexico, all of which are worked for silver. They are rarely noted in the western part of the United States. Much remains to be learned about their relation to the metallization. In general, it seems certain that the zeolites were deposited later than the other minerals, probably not, however, by descending waters, but rather by remaining stagnant parts of the original vein-forming solutions. A concentration of silver ores often accompanied their development. Calcite, quartz, barite, and fluorite are the principal gangue minerals in the typical localities; the presence of antimony and quicksilver is often mentioned.

The renowned silver mines of Kongsberg, in southern Norway, which have been worked for several hundred years and still remain productive, have been described by Vogt.[1] The deposits are narrow veins in gneiss and mica schist, often breaking up when entering amphibolite. Along certain lines following the schistosity the rocks contain disseminated sulphides, mainly pyrite and pyrrhotite, and the veins become enriched where crossing these "fahlbands," probably on account of their precipitating influence. The mines have been worked to a depth of 3,000 feet. Quartz, chlorite, and axinite crystallize next to the walls, but the prevailing gangue is calcite with some barite and fluorite, rarely adularia and albite. Zeolites also occur and are among the latest gangue minerals, prehnite, stilbite, harmotome, and laumontite being among those identified. The prin-

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[1] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1899, pp. 113–123; 177–181.

cipal ore mineral is native silver, mostly in wire form; this is believed to be derived from the more scarce argentite by a process of enrichment. Less prominent are ruby silver, stephanite, pyrite, pyrrhotite, arsenopyrite, chalcopyrite, zinc blende, and galena, the latter poor in silver. A certain part of the silver is believed by Vogt to result from primary deposition. The native silver contains quicksilver. Anthracite is also one of the gangue minerals deposited during the early stages.

Vogt supposes the native silver to be derived from argentite and proustite as follows:

$$Ag_2S+O_2=2Ag+SO_2.$$
$$Ag_2S+H_2O=2Ag+H_2S+O.$$
$$Ag_3AsS_3+3H_2O=3Ag+As+3H_2S+3O.$$

The list of minerals given shows clearly that the veins have had a complicated history, beginning with the deposition of high-temperature minerals like axinite and ending with that of minerals like zeolites, probably formed at about 100° C. This history has evidently not yet been followed in detail; it is stated by Vogt that axinite crystallized together with the zeolites, but this seems a curious association. The presence of free oxygen at great depths might also well be questioned.

The other notable occurrence is at Andreasberg, in the Harz Mountains, best described by A. Bergeat.[1] The veins at Andreasberg are simple filled fissures, at most 0.5 meter thick, chiefly in Silurian clay slates and quartzites. They appear not far from the intrusive mass of the "Brocken" granite. The veins are included between two great divergent dislocations, forming impermeable walls, against which the silver veins split and cease. The mines, which attained a depth of 2,700 feet, are now closed.

In general the veins carry argentiferous galena and tetrahedrite; sometimes they yield large druses full of rich silver ores, calcite, and zeolites. Bergeat distinguishes five phases:

1. In crevices near the veins, in part also in the fissures themselves, are garnet, epidote, axinite, and albite.
2. Earliest bituminous calcite with simple antimonides, arsenides, and sulphides: Niccolite, smaltite, löllingite, breithauptite (NiSb), zinc blende, galena, pyrite, pyrrhotite, chalcopyrite.
3. Tetrahedrite with quartz (replacing calcite) and fluorite,

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[1] Stelzner and Bergeat, Die Erzlagerstätten, vol. 2, 1906, pp. 718–720

chalcopyrite, galena, zinc blende; native silver and millerite, encrusting tetrahedrite.

4. Sulphantimonides and sulpharsenides: Pyrargyrite, proustite, miargyrite, polybasite, stibnite, argentite (in part from pyrargyrite), fluorite.

5. Native silver, realgar, calcite, apophyllite, analcite, chabazite, heulandite, brewsterite, harmotome, stilbite, natrolite; also fluorite and chalcopyrite.

This clear exposition of paragenesis indicates a long epoch of deposition with gradually diminishing temperature and emphasizes the connection of the native silver with zeolitization.

Zeolites occur in a number of Chilean silver veins, particularly at Arqueros and Rodaito, in association with native silver, calcite, barite, and silver amalgam. The only occurrence known in the United States is at the South Republic mine, Republic, Washington, where the gold selenide veins in andesite are filled by closely banded, extremely fine grained quartz (p. 527). At the mine mentioned the vein has in part been dissolved and replaced by a loose aggregate of calcite and laumontite much richer in silver than the original quartz.

The occurrence of native silver in the zeolitic copper and silver deposits of the Lake Superior region is described on page 437. In short, zeolitization, probably effected by warm or tepid waters, seems particularly adapted to the concentration of silver and the deposition of both native silver and rich silver minerals.

The Silver-Bearing Cobalt-Nickel Veins of Saxony.—In different parts of the world occur narrow veins with calcite or barite gangue and arsenides or sulphides of cobalt and nickel; the cobalt and nickel minerals contain silver, and this metal is often separated as seams of native silver enclosed in the older metallic minerals. At the present time it seems probable that the silver is primary and that it was deposited by the same solutions from which the earlier arsenides were formed.

The cobalt veins of Annaberg,[1] in Saxony, appear in gneiss intruded by dikes of granitic and lamprophyric character; they are younger than the veins in the same region carrying cassiterite and those yielding pyritic ores with galena. The gangue minerals are barite, calcite, fluorite, quartz, and dolomitic car-

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[1] H. Müller, Die Erzgänge des Annaberger Bergrevieres, Geol. Landesanstalt, Leipzig, 1894.

bonates. The principal ore minerals are chloanthite, smaltite, bismuthinite, also rich silver minerals such as argentite, pyrargyrite, and native silver; the latter are distinctly later than the primary nickel-cobalt-bismuth ores.

Most of the rich silver ores appear where the veins intersect certain flat crushed zones in the gneiss, which contain carbonaceous material and finely divided sulphides like pyrite and chalcopyrite. The greatest depth attained was 1,400 feet.

The veins of Schneeberg, in Saxony,[1] are contained in contact-metamorphic clay slates and tend to impoverishment in the underlying granite. The primary gangue consisted of calcite, ankerite, barite, and fluorspar, but these minerals are now largely replaced by hackly and platy fine-grained quartz by a process similar to that to which many later Tertiary gold-silver veins have been subjected. This is thought to be the only locality where such a replacement has been carried on in veins of more deep-seated deposition. The ore minerals are smaltite, chloanthite, niccolite, bismuthite, and native bismuth. Native silver and rich silver minerals are subordinate in the silicified veins, but appear in the primary barytic veins. From this it is perhaps permissible to draw the conclusion that the silicification has been accompanied by solution and removal of silver. The process was evidently not effected by surface waters, but rather by ascending siliceous solutions.

Uranium ores, mainly uraninite or pitch blende, are found at Schneeberg and, more abundantly, in the somewhat similar veins at Joachimsthal,[2] Bohemia. The geological relations at the two places are similar. At both places the cobalt and nickel minerals are the older and the rich silver minerals the younger. Between them in point of age lie the uranium ores.

The Silver-Bearing Cobalt-Nickel Veins of Ontario, Canada.[3] —At a number of localities in Ontario silver-bearing veins have been found. Many of them appear to be connected genetically with igneous rocks of Keweenawan age, and the occurrence of

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[1] K. Dalmer, E. Köhler, and H. Müller, Section Schneeberg, *Geol. Spez. Karte Sachsen,* 1883.

[2] J. Step and F. Becke, K. Akad. Wis. Wien Sitzungsber., vol. 113, 1904, pp. 585–618; also E. S. Bastin and J. M. Hill, *Prof. Paper* 94, U. S. Geol. Survey, 1917, p. 122.

[3] E. D. Ingall, Report on mine and mining on Lake Superior (Silver Islet), *Ann. Rept.* Canada Geol. Survey, pt. H, 1887.

Willett G. Miller, Cobalt-nickel arsenides and silver deposits of Temnative silver in the copper mines working the amygdaloids of Michigan may be recalled.

The veins of Silver Islet and those at points southwest of Port Arthur, on the north shore of Lake Superior, have been known for many years. The Silver Islet vein, at one time a heavy producer, intersects greenstone and contains, in a gangue of calcite, ankerite, and quartz, native silver, argentite, tetrahedrite, galena, zinc blende, pyrite, and some cobalt and nickel minerals. Graphite, it is stated, occurs also in the vein.

In 1903 the silver veins of Cobalt, Ontario, were discovered and they soon became extraordinarily productive. Up to 1917, inclusive, the total production had reached about 275,000,000 ounces, and the output in that year was 19,250,000 ounces. The ore is extremely rich, its tenor often reaching several thousand ounces per ton. Some of the veins contained slabs of native silver of great size. One specimen, now in the Parliament building at Toronto, is 5 feet long and weighs 1,640 pounds; it contains 9,715 troy ounces of silver. The La Rose vein, in a horizontal distance of 100 feet and above the 60-foot level, yielded 532 tons, which contained 658,000 ounces of silver. So-called low-grade ores average about 200 ounces per ton, while that which is concentrated before shipping contains about 30 ounces per ton. Owing to the complex character of the ore the smelting charges have been very high. Lately the cyanide, amalgamation and flotation processes have been adopted with much success. Besides silver, cobalt, nickel and arsenic are recovered.

In the last few years similar cobalt-silver veins, productive in part, have been discovered in other portions of Ontario, particularly at Gowganda and South Lorrain.

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iskaming, *Fourteenth Ann. Rept.*, Ontario Bur. Mines, pt. 2, 1905. *Idem. Sixteenth Ann. Rept.*, pt. 2, 1907; *Nineteenth Ann. Rept.*, pt. 2, 1913.

H. V. Ellsworth, A study of certain minerals from Cobalt, *Twenty-fifth Ann. Rept.*, *idem*, 1916, pp. 200–243.

A. A. Cole, The mining industry of part of Ontario served by T. and N. O. Ry., Annual publication.

R. E. Hore, Origin of Cobalt silver ores, *Trans.* Canadian Min. Inst., vol. 2, 1908, pp. 275–286.

S. F. Emmons, Cobalt district, Ontario, *Min. and Sci. Press*, March 18, 1911, reprinted in "Types of Ore Deposits," San Francisco, 1911.

W. Campbell and C. W. Knight, A microscopic examination of the cobalt-nickel arsenides, *Econ. Geol.*, vol. 1, 1906, pp. 767–776.

E. S. Bastin, Significant mineralogical relation in silver ores of Cobalt, *Econ. Geol.*, vol. 12, 1917, pp. 219–236.

The geology of the Cobalt region is summarized by W. G. Miller as follows: The oldest rocks, known as the Keewatin series, are basic volcanic rocks, greenstones, and schists, with more or less cherty or jaspery sediments. On the eroded Keewatin were deposited the conglomerate and graywacke (arkose sandstone) of Temiskaming (Huronian) age. A thickness of 300 feet of these gently dipping strata is exposed at Cobalt.

After the deposition of the Huronian beds an irruption of diabase took place, assuming the form of sills from 100 to 500 feet thick. The veins were formed after the intrusion of this diabase, which is regarded by some authors as of Keweenawan age; the fractures occupied by the veins are believed by some to have been caused by cooling or shrinking of the diabase, as well as the Keewatin and Huronian. Fig. 218 illustrates the geological relations and the subsequent erosion of the region to the present surface. The veins in the Huronian conglomerate have been the most productive, but ores occur also in the Keewatin and in diabase.

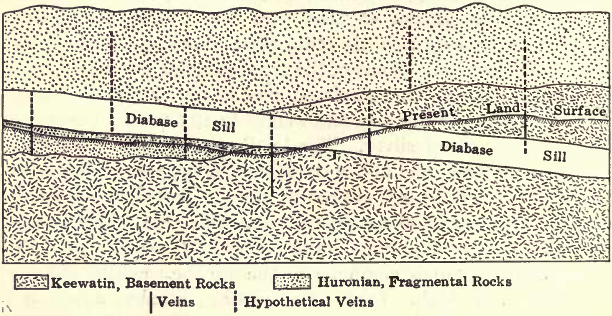

FIG. 218.—Geological section of Cobalt district, Ontario, showing also probable geological relations before erosion had produced the present land surface. *After W. G. Miller.*

Almost the whole production has been derived from veins in the lower wall of the sill, the veins once cutting through the upper or hanging wall having been mostly removed by erosion. Few of the veins have been followed more than 500 feet horizontally and very little ore has been taken out below the 500-foot level. Some ore has, however, been found at a depth of 1000 feet.

The greatest depth reached in the district is 1600 feet (1917). About 100 veins are known in the district.

In the narrow veins, which are seldom more than a few inches in width, the filling is often "frozen to the walls." (Fig. 219.) In some places the veins form a network of stringers. The silver, cobalt, and nickel are very irregularly distributed in the veins; the adjoining country rock often contains carbonates of dolomitic character and small specks of the various minerals, including native silver. The gangue is calcite or a calcium-magnesium carbonate; graphite is reported; quartz occurs sparingly. The ore minerals are abundant and consist of native silver, native bismuth, niccolite (NiAs), breithauptite (NiSb), smaltite ($CoAs_2$), chloanthite ($NiAs_2$), argentite, millerite, arsenopyrite, cobaltite (CoAsS), dyscrasite (Ag_6Sb), pyrargyrite, stephanite, polybasite, proustite, and tetrahedrite.[1] Pyrite is rare. Smaltite and niccolite are most common. The minerals have a marked

FIG. 219.—Photograph of vein of arsenides with inclusions of greywacke, Cobalt district. *After W. L. Whitehead.*

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[1] The silver and the dyscrasite contains in places a little mercury, rising in exceptional cases to 5 per cent. It is believed to be present as silver amalgam. G. H. Clevenger, *Econ. Geol.*, vol. 10, 1915, p. 770.

tendency to concentric arrangement and dendritic replacements of calcite by smaltite and silver are often seen. Coatings of green nickel arsenate and pink cobalt arsenate mark the outcrops.

Carload lots, containing from 100 to 6,000 ounces per ton, averaged 6 per cent. Co, 3.66 per cent. Ni, and 27 per cent. As.

An analysis by A. R. Ledoux of two carloads shipped to the smelter is as follows:

SiO_2	3.34	Bi	trace
Fe	1.78	Ag	5.31
Al_2O_3	0.27	Sb	1.46
CaO	5.85	As	42.46
MgO	4.63	CO_2	9.26
Cu	0.09	Cl	0.08
Ni	13.87	S	1.89
Co	8.63		

The majority of the authors have expressed the belief that the ore deposition is genetically connected with the intrusion of the diabase sills which are probably of Keewenawan age[1] and it is also generally recognized that the native silver is later than the cobalt and nickel arsenides. Some veins rich in arsenides are poor in silver. The order of succession was first roughly established by Campbell and Knight. Ellsworth concludes that the mineralizing solutions were at first rich in arsenic and deposited smaltite and chloanthite. Later the monarsenides, niccolite and breithauptite were formed. Finally sulphur became prominent and the sulpharsenides such as arsenopyrite and cobaltite were precipitated. Calcite was then deposited and, after a period of fracturing, solutions which may have contained sulphate or carbonate of silver began to deposit native silver mainly by replacement of the diarsenides and monarsenides while the sulpharsenides and pyrite were inert.[2]

Native bismuth, argentite, proustite, etc., were deposited with the silver. The process of deposition of the various arsenides was practically continuous but the silver distinctly marks another epoch. Whether this is a case of enrichment by descending solutions which have leached the upper, now eroded, parts of the

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[1] C. K. Leith has pointed out the remarkable association of silver deposition in the Lake Superior region with the Keewenawan intrusions.

[2] Chase Palmer, Diarsenides as silver precipitants, *Econ. Geol.*, vol. 12, 1917, pp. 207–218.

vein, or whether it is a separate phase of the primary deposition is not yet fully established. Emmons, Van Hise, and Bastin have held the former, whereas some observers like W. G. Miller, A. R. Whitman and W. L. Whitehead (unpublished manuscripts) are inclined to the other view.

While the question is a difficult one it does not seem improbable that the silver may have been deposited by the last ascending solutions of magmatic origin.

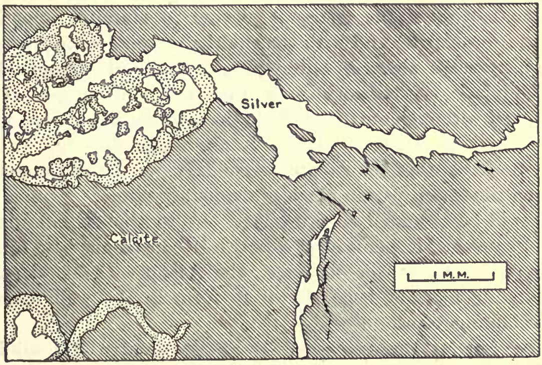

FIG. 220.—Replacement veinlet of silver traversing ferruginous calcite and smaltite (with niccolite). Arsenides are dotted. *After E. S. Bastin.*

QUARTZ-ADULARIA-ZEOLITE VEINS (ALPINE TYPE)

Occurrence and Mineral Association.—The so-called Alpine veins have little or no importance as a source of metalliferous ores, but many of them contain beautifully developed crystals, represented in all large mineral collections, and are prospected and worked to some extent for such specimens. They occur in the crystalline rocks in Tyrol, Switzerland, and the French Alps. There are several types. One class found in the Zillerthal, in Tyrol, contains pyrite and galena with quartz, adularia, albite, epidote, calcite, prehnite, desmine, and laumontite. Another, in the French Alps, yields axinite, titanite, ilmenite, and many other minerals which indicate deposition at high temperature.

The veins in Switzerland have been studied in detail by J.

Königsberger,[1] whose investigations have shed much light on the conditions under which these beautiful minerals have been formed.

The "veins" are approximately horizontal filled crevices in a biotite gneiss; they are really local openings in joints, which can be followed for considerable distances and which lie perpendicular to the schistosity of the rock. These crevices are surrounded by zones of altered rock not more than double the width of the opening; they usually contain open cavities into which the crystals project.

The biotite of the gneiss is altered to chlorite; sometimes also to specularite, and with the chlorite are epidote and quartz; the plagioclase alters to sillimanite, kaolin, and epidote; albite and adularia crystallize in vugs; quartz and orthoclase are not attacked, although next to the vein the orthoclase is often covered by secondary adularia. While Königsberger does not say so, it is probable that the kaolin is secondary, for the veins have for a long time been within reach of oxidizing waters.

It will be observed that the type of alteration is not that of ordinary veins; it seems most closely related to some veins formed near the surface—for instance, the Cripple Creek deposits. Sillimanite has not been shown to form in the wall rocks of metalliferous veins.

The general succession is: (1) Smoky quartz and adularia (oldest); (2) calcite; (3) zeolites. The zeolites are not always noted in descriptions of Alpine veins, because they have often been leached out or softened by oxidizing waters. The long list of other minerals includes fluorite, chlorite (often dusting the faces of quartz and adularia), apatite, albite, stilbite, heulandite, apophyllite, laumontite, chabazite, specularite, and rarely galena, chalcopyrite, and molybdenite.

Origin.—Königsberger concludes that the minerals were derived chiefly from the rock itself. The minerals were deposited by cooling of hot solutions containing carbon dioxide. The crystallization in the crevice took place at temperatures of 290° to 130°C. It is not improbable that the Alpine veins were produced by intrusive after-effects at various temperatures.

Some of the quartz crystals contain large fluid inclusions; an analysis of the liquid gives the following result. The total solids amount to 10 per cent., a rather concentrated solution.

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[1] J. Königsberger, Die Minerallagerstätten im Biotitprotogin des Aar massives, *Neues Jahrb., Beil. Bd.*, vol. 14, 1901, pp. 43–49.

COMPOSITION OF FLUID INCLUSION IN QUARTZ

H_2O	85.0
CO_2	5.0
Na	2.5
(K, Li)	1.5
CO_3	3.5
Cl	1.5
SO_4	0.7
	99.7

The dissolved salts then consist of alkaline carbonates, chlorides, and sulphates in the order named.

As the smoky quartz is bleached at 300° C. it is assumed that it could not have been formed above this temperature. The bubbles of the inclusions disappear at 225° C. Because the liquid must have filled the cavity at the time of formation and because the pressure did not differ much from that of saturated water gas, the temperature of formation is about the same as that at which the bubbles disappear.

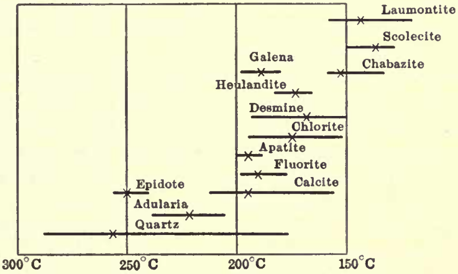

FIG. 221.—Order of crystallization in the zeolitic veins of Aar, Switzerland. *After J. Königsberger.*

The red color of the fluorite disappears at 175° C., independent of the surrounding medium and pressure. The zeolites were probably formed at about 130° C. Königsberger's view as to the order of crystallization is expressed in Fig. 221.

Analogous minerals are found in the miarolitic cavities in certain granites;[1] in these occurrences the succession, beginning

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[1] A. Schwantke, Drusen Mineralien des Striegauer Granits, *Inaug. Dissert.*, Leipzig, 1890.

with the earliest, is (1) adularia, albite, quartz, epidote; (2) fluorite, apatite, calcite, chlorite, epidote; and (3) zeolites.

THE COPPER VEINS

The copper-bearing veins in which filling has been the principal process do not have the importance of the corresponding class of gold quartz veins. Only when replacement has converted wider bodies of rock into ore and when enrichment by descending surface waters have acted upon the primary ore do we find large and productive members in this class.

Many of the great copper deposits are mainly pyritic replacements of igneous or sedimentary rocks and are described on the following pages or with the high-temperature deposits in Chapter XXVI. Still others derive their importance from the accumulation of secondary chalcocite, either in wide replacement veins or in broad mineralized zones. Many of these in fact also belong to the high-temperature deposits as far as the poor, primary mineralization is concerned.

Chalcopyrite-Quartz Veins.—Simple veins containing mainly chalcopyrite, pyrite and quartz, with some bornite, more rarely tetrahedrite and siderite, are common enough in many districts, but are rarely of great importance unless also containing gold and silver. The extensive group of tourmaline-bearing copper veins belongs to the high-temperature deposits.

J. B. Umpleby[1] describes the Lost Packer vein, in central Idaho, which cuts through mica schist and which, in a gangue of quartz and siderite, contains chalcopyrite and some pyrrhotite and pyrite. The chalcopyrite and the quartz contain, in the ore shoots, about 3 ounces of gold per ton. The vein is intersected by dikes and is probably a high-temperature deposit.

Bornite-Quartz Veins.—Quartz veins containing bornite are not uncommon but are rarely of great economic importance. Veins of this kind occur in the Virgilina district,[2] Virginia and North Carolina, where F. B. Laney noted interesting intergrowths of bornite and chalcocite. The deposits are, however, lenticular veins and probably were formed at high temperatures.

Pyrite-Enargite Veins.—The enargite-bearing veins constitute a less common yet in places a most important class.

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[1] *Bull.* 530, U. S. Geol. Survey, 1913, pp. 70–73.

[2] W. H. Weed, *Bull.* 455, U. S. Geol. Survey, 1911, pp. 67–93.

F. B. Laney, *Bull.* 21, North Carolina Geol. and Econ. Survey, 1910; *Bull.* 14, Virginia Geol. Survey, 1917.

Enargite is, on the whole, a rare mineral and rather favors the deposits formed relatively near the surface. At Mancayan, in Luzon, in the Philippine Islands, enargite is found in a large replacement deposit in "quartz porphyry" and "andesite," formerly worked on a large scale and containing, in a quartz gangue, tetrahedrite, bornite, and enargite (luzonite). Another deposit is that of Bor,[1] in Serbia, which is also an irregular quartzoze replacement in a rock that is called andesite but is really an intrusive porphyry with holocrystalline groundmass. The principal minerals are pyrite, enargite, quartz, barite, and secondary covellite. Finally should be mentioned the occurrence in the Sierra Famatina,[2] Argentina, where veins carrying quartz, enargite, and famatinite (Cu_3SbS_4) break through clay slate intruded by granitic rocks and "dacite."

The most prominent examples of this type are found in the Butte district[3] (Figs. 46 and 47), Montana, now the most productive copper camp in the world. The veins are contained in sericitized quartz monzonite and are formed in part by replacement of the intrusive rock. The primary ores are pyrite, enargite (Cu_3AsS_4), and perhaps chalcocite, with smaller amounts of bornite and sphalerite. On account of the importance of secondary processes in their formation the description of the Butte copper veins is placed in Chapter XXXI. Similar veins occur at Cerro de Pasco, and Morococha, in Peru, and at Chuquicamata in Chile.

THE PYRITIC REPLACEMENT DEPOSITS

While pyrite is a persistent mineral, crystallizing within a wide range of temperatures, it is easily apparent that the deposits containing large masses of pyrite have not been formed close to the surface, but rather at considerable depths and at temperatures well above 100°C. The deposits of this type are mainly confined to regions that have been deeply eroded since the deposits were formed and many of them show a mineral association indicating high temperature.

In many text-books the pyritic deposits are treated as a distinct class, and are assumed to have a similar origin. We know

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[1] M. Lazarevic, *Zeitschr. prakt. Geol.*, 1912, pp. 337–370.

[2] A. W. Stelzner, Beiträge zur Geologie der Argentinischen Republik, pt. 1, 1885, p. 228.

[3] W. H. Weed, Geology and ore deposits of the Butte district, Montana, *Prof. Paper* 74, U. S. Geol. Survey, 1912. See also footnotes, p. 862.

now that they comprise deposits of widely differing origin and history.

In a broad way we may distinguish (1) those associated with silicates such as amphibole, pyroxene, epidote, tourmaline, and garnet, the iron sulphide being in part present as pyrrhotite, and (2) those associated with calcite, barite, and quartz as gangue minerals. The deposits of the first class were undoubtedly formed at considerably higher temperatures than those of the second and, in general, probably also at greater depth.

Class 1 comprises (A) some deposits of purely magmatic origin like those of Sudbury, Ontario; (B) a large division, of contact-metamorphic type, like the deposits of Ducktown, Tennessee; Granby, British Columbia; and the Highland Boy mine at Bingham, Utah; (C) a third division, difficult to interpret but thought by many to represent a phase of igneous injection; this may not be firmly established, but their close connection with igneous rocks can hardly be questioned; among these are the deposits at Vigsnäs, Sulitelma, and Röros in Norway, Fahlun and Bersbo in Sweden, and Bodenmais in Bavaria. In each of these three divisions the deposit may have been subjected to dynamic metamorphism, with the attendant development of amphibole and garnet and of schistose structure. Many of these metamorphosed deposits have had a complicated history and are among the most difficult to interpret and classify.

Class 2 is also connected with the eruption of igneous rocks, but the high-temperature minerals are absent; the deposits of Rammelsberg in the Harz, Germany, of Mount Lyell, Tasmania, of Rio Tinto, Spain, and of Shasta County, California, may serve as examples. The deposit at Kyschtim,[1] in the Ural Mountains, and that at Tyee,[2] on Vancouver Island, also appear to belong in this class.

Dynamic metamorphism may produce remarkable changes in structure and mineral association. Replacement of various rocks by pyrite has played an important part; sometimes this process takes place in shattered zones or proceeds from fissures, or again it may be caused by solutions permeating heated limestone masses without fractures at igneous contacts.

The ores, while consisting mainly of pyrite or pyrrhotite, derive

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[1] A. W. Stickney, *Econ. Geol.*, vol. 10, 1915, pp. 593–633.

[2] C. H. Clapp, *Mem.* 13, Canada Geol. Survey, 1912, pp. 180–187.
V. Dolmage, *Econ. Geol.*, vol. 11, 1916, pp. 390–394.

their value from a small percentage of chalcopyrite; there are usually minute quantities of gold and silver, and frequently also zinc blende and a little galena; other sulphides are rare. In all the deposits mentioned, except those at Rammelsberg, their connection with rhyolite porphyries, alaskite porphyry, or keratophyre can be established, and deposition by hydrothermal replacement at moderate depth and temperature seems the most reasonable explanation.

The deposits of Class 1 are described in Chapters XXVII and XXVIII. Some pyritic deposits of Class 2 will be briefly characterized in the following paragraphs.

Copper Deposits of Shasta County, California[1]

Copper deposits which have been actively mined and smelted since 1895 are found in a number of districts in Shasta County, California; among the more prominent mines are the Iron Mountain, Bully Hill, Mammoth, and Balaklala. The production of copper in 1917 was 26,700,000 pounds.

The sedimentary rocks consist of Devonian and Carboniferous closely folded slates and contain intrusions of a highly siliceous and sodic alaskite porphyry, which is the country rock of almost all the important copper deposits. Somewhat later than the alaskite porphyry, but belonging to the same (early Cretaceous) period of intrusion, is a quartz diorite, probably equivalent to the granodiorite of the Sierra Nevada. No copper deposits occur in the quartz diorite, but it contains workable gold-bearing quartz veins. Deep erosion has taken place since the period of intrusion; Graton estimates the depth of rocks removed as not less than 5,000 or 6,000 feet. The rocks have been subjected to some shearing and brecciation, but little extensive dynamo-metamorphism, since the intrusion.

The copper deposits were formed during the interval between the two epochs of intrusion. The ore-bodies are large, irregular tabular masses of pyrite with some chalcopyrite (Fig. 222);

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[1] J. S. Diller, *Redding Folio* 138, U. S. Geol. Survey, 1906.

L. C. Graton, The occurrence of copper in Shasta County, Cal., *Bull.* 430, U. S. Geol. Survey, 1910, pp. 71–111.

A. C. Boyle, The geology and ore deposits of the Bully Hill mining district, *Trans.* Am. Inst. Min. Eng., vol. 48, 1915, pp. 67–117.

single ore masses have dimensions of 1,200 feet in length, 300 feet in width, and nearly 300 feet in thickness, and some of them contain many million tons of ore; the Iron Mountain mass before a great part of it was converted to gossan probably contained 20,000,000 tons of ore, exclusive of the large amount which has been removed by erosion. Many of the bodies lie flat and are easily accessible by tunnels.

The ores contain chiefly pyrite with about 3 per cent. of copper as chalcopyrite, and as much as $2 per ton in gold and silver, about equally divided between the two metals. Zinc blende is present in varying amounts, and the ore contains a little bismuth, arsenic, and selenium. The gangue minerals include quartz, calcite, barite, gypsum and anhydrite; the succession is in general pyrite (oldest), zinc blende, chalcopyrite, quartz, and barite. There is a deep oxidized zone with sulphide enrichment.

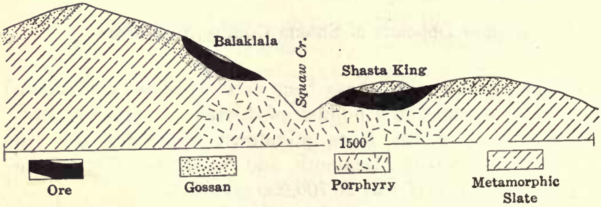

Fig. 222.—Cross-section of ore-bodies at Balaklala, California. *After W. H. Weed.*

The alaskite porphyry near the ore-bodies is more or less altered and contains sericite (probably also paragonite), secondary quartz, chlorite, pyrite, carbonates, and epidote. Cogent evidence is cited by Graton that the pyritic ores are replacements of the surrounding porphyry in sheared and brecciated zones. This replacement is believed to be due to hot solutions emanating from the cooling alaskite porphyry. The action of surface waters in the ore deposition is probably negligible, for at that time the alaskite porphyry was everywhere covered with a blanket of impermeable shales.

The Pyritic Deposit of Mount Lyell, Tasmania[1]

At Mount Lyell, on the west coast of Tasmania, is one of the large copper deposits of the world. According to Gregory the ore-bodies are contained in sericite schists (probably with paragonite), which are dynamo-metamorphic forms of perhaps Paleozoic acidic porphyries. Intrusive in these schists are considerable masses of igneous rocks which are termed "porphyrites," but of which no analyses are available. Probably of later age than the complexes just mentioned are Devonian conglomerates.

The main ore-bodies are lenticular and lie in part in the sericite schist and in part at its contact with the conglomerates which have been brought against the schist by faulting. The largest ore-body is that of the Mount Lyell mine; this is several hundred feet long and 200 feet wide, but appears to be limited in depth by a fault.

The ore consists mainly of pyrite with but little gangue of quartz and barite. It also contains 5 to 6 per cent. of copper in the form of chalcopyrite, more rarely bornite and enargite; an analysis of the ore given by Gregory is as follows:

Fe	40.30
SiO_2	4.42
$BaSO_4$	2.50
Cu	2.35
Al_2O_3	2.04

The ore yields very little gold and 1.33 ounces of silver to the ton. The output of copper was 6,500 tons in 1918. Pyrite is the oldest mineral; it was followed by chalcopyrite, bornite and enargite.

Enrichment near the surface and in the footwall of the deposit has added much to the wealth of the property. The croppings at one place contained quartz, barite, hematite, and about 15 ounces of silver and $15 in gold per ton. The presence of hematite in the croppings would suggest that at some time, during the weathering of the deposits, the climate was warmer than now. The secondary sulphides consisted of chalcopyrite, bornite, and tennantite, with stromeyerite.

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[1] J. W. Gregory, The Mount Lyell mining field, Tasmania, *Trans.* Aust. Inst. Min. Eng., Melbourne, vol. 10, 1905, pp. 26–197.

C. G. Gilbert and J. E. Pogue, The Mount Lyell copper district of Tasmania, *Proc.* U. S. Nat. Mus., vol. 45, 1913, pp. 609–625.

Gregory as well as Gilbert and Pogue find that the ore minerals replace the schists and the latter suggest a relationship between the "porphyrites" and the pyritic deposits.

The Pyritic Deposits of Rio Tinto, Spain[1]

General Features.—The pyritic ore-bodies of the southern Spanish province of Huelva, more generally known as the deposits of Rio Tinto, are probably the greatest in the world and have been mined since Phoenician and Roman times. The deposits are in the main lenticular; there are at least 50 of these pyritic lenses, whose length varies from 1,200 to 6,500 feet, while the width, in general proportional to the length, reaches a maximum of 250 feet and the depth ranges from 500 to 1,800 feet. The vertical range of deposition, according to Finlayson, probably in no case exceeded 3,300 feet, and few of the deposits attain a depth of 1,000 feet. In the slates the deposits often taper downward to a point, while in the porphyry a flat or rounded lower surface is not uncommonly observed. On the whole they appear to lie conformably between slates and porphyry or in either porphyry or slate.

The production of these deposits has always been large, but appears now to be diminishing; in 1917 it was 40,000 metric tons of copper, all sources considered. Besides the regular copper ore with more than 2 per cent. Cu, large quantities of pyrite, poor in copper, are shipped for sulphuric acid manufacture. A part of the copper is recovered as a precipitated cement or a sulphate.

Geological Formations.—The rocks consist of (1) a uniform series of folded and compressed clay slates and graywacke, striking east and west and believed to be of Devonian and Carbonifer-

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[1] A. Moncrieff Finlayson, The pyritic deposits of Huelva, Spain, *Econ. Geol.*, vol. 5, 1910, pp. 357–372; 403–437.

Bruno Wetzig, Beiträge zur Kenntniss der Huelvaner Kieslagerstätten. *Zeitschr. prakt. Geol.*, vol. 14, 1905, p. 173.

H. Preiswerk, Die Erzlagerstätten von Cala, etc., *Idem*, vol. 12, p. 225.

F. Klockmann, Ueber das Auftreten, etc., der Südspanischen Kieslagerstätten, *Idem*, vol. 10, 1901, p. 113; also vol. 3, 1894, p. 35.

J. H. L. Vogt, Das Huelva Kiesfeld, etc., *Idem*, vol. 7, 1898, p. 241.

L. de Launay, Mémoire sur l'industrie, etc., dans la région d'Huelva. *Ann. des Mines*, Paris, ser. 7, vol. 16, p. 407.

J. Gonzalo y Tarin, Descripción, etc., de la provincia de Huelva, *Mem.* Com. Mapa Geol. España, Madrid, vol. 1. 1886.

ous age; (2) granites and granodiorites intrusive into the Carboniferous rocks, north of the district; (3) several varieties of porphyry, including rhyolite porphyry and keratophyre, some varieties with granophyric structure; in places the porphyry is affected by shearing and schistosity; (4) basic dikes and sills, mainly of diabase, but including also camptonites and diorites. Some authors, including Klockmann, consider the porphyries as effusive, but the arguments of Finlayson show quite conclusively that they are intrusive masses, occurring in belts and lenses throughout the field (Fig. 223). The basic rocks cut both granites and porphyries.

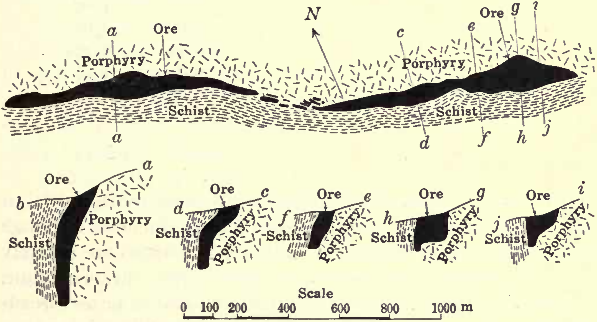

FIG. 223.—Cross-section of ore-bodies, Rio Tinto, Spain. *After Gonzales y Tarin.*

Displacements of considerable throw occur along many ore-bodies. According to Finlayson, the last event in this series of igneous and dynamic disturbances consisted in the development of the mineral deposits.

The lodes that occur in the slate are in the main conformable with the bedding, but the ore sometimes, according to Finlayson, cuts across it; the lode walls are usually well defined and smooth; the deposits occur, as a rule, along contacts or other lines of weakness and crushing. According to the same author the adjoining rock shows effects of hydrothermal action in marked degree, the porphyry being transformed into an aggregate of chlorite, sericite, quartz, carbonates, and pyrite. Analyses show extremely well marked carbonatization and sericitization, entirely similar to the alteration occurring in the California type of gold-quartz veins, and undoubtedly of hydrothermal origin. The alteration is illustrated by the following analyses, quoted from Finlayson:

ANALYSES OF FRESH AND ALTERED PORPHYRY FROM THE SAN DIONISIO MINE, RIO TINTO, SPAIN

	Fresh.	Altered.
SiO_2	76.21	70.68
Al_2O_3	12.66	11.45
Fe_2O_3	2.98	1.31
FeO	1.46	0.72
MnO	0.08	0.05
CaO	1.15	2.28
MgO	0.10	0.17
K_2O	3.27	4.85
Na_2O	1.64	0.65
H_2O+	0.18	0.23
H_2O-	0.35	1.41
CO_2	0.09	5.08
FeS_2		1.27
	100.17	100.15

The Ores.—The ores consist of almost massive pyrite, with but a small amount of quartz and few other gangue minerals, although barite occurs in some localities. Banded or pressed ore is rarely seen. The primary ore carries from 48 to 50 per cent. of sulphur. Chalcopyrite is present in minute scattered grains, or as threads and strings traversing the granular pyrite and filling interstices in the ore. Blende and galena are present in small amounts, and there are traces of bismuth, selenium, and tellurium. Arsenic varies from 0.25 to 1 per cent. The order of succession is pyrite (oldest), chalcopyrite, blende, galena.

Especially interesting are the changes in the ore produced by weathering. The outcrops are gossans of hematite carrying 10 to 15 per cent. of silica and alumina and little or no copper. The average depth of this gossan is 100 feet. The lower limit of the gossan is well defined, and the line of contact between it and the sulphide ore is sometimes marked by a thin earthy material, which, as described by Vogt, is rich in gold and silver. The top portion of the sulphide zone, for a thickness of 3 feet or more, is composed of leached pyrite with traces of copper (Finlayson). Below this commences the zone of enriched sulphides, in which the ore assays from 3 to 12 per cent. of copper. This enrichment is effected by deposits of both chalcopyrite and chalcocite, and its influence may be traced to depths of 300 feet below the surface. In the San Dionisio lode enrichment was noted even at a depth of 1,000 feet, indicating, according to Finlayson, that the secondary changes extend far down into what is usually regarded as primary ore. Wetzig states that in the Cabeza de Pasto mine the ore at the 40-meter level contained 3.5 per cent. of copper, at the 60-meter level 3 per cent., and at the 80-meter level 2 per cent. The tenor of the primary ore ranges from 0.5 to 2 per cent. of copper.

Genesis.—The origin of these deposits has been the subject of long discussion among geologists. The earlier geologists believed in a sedimentary origin, and this view is still held by some. Its principal defender is F. Klockmann, who regards the pyrite bodies as concretionary deposits in mud, impregnated by copper from the supposedly effusive lavas. Gonzalo y Tarin, as well as De Launay, held them to be veins or lodes deposited in open cavities by ascending solutions. Later, Vogt considered the deposits to be of pneumatolytic nature, formed as an after-effect of the extrusions of porphyry. In the latest contribution by Finlayson the metasomatic character of the deposits, which were formed by the hydrothermal replacement of crushed and sheared zones, appears to be firmly established. He believes, however, that the deposits were formed after the intrusion of the basic dikes and sills, which are considerably later than the porphyry, and thinks that the concentration of the ores was in the first place due to a process of magmatic concentration of sulphides, accompanying the differentiation of the series of intrusive rocks and dependent, with the latter, on the Hercynian tectonic movements.

Expression of opinion without field acquaintance may have little value, but it seems to me that it has not been definitely shown that the deposits are later than the basic intrusions. They are clearly of hydrothermal origin, as shown by the character of alteration, and the replacement origin seems definitely proved. There appears to be good reason for Vogt's view that the hydrothermal processes of deposition followed the intrusion of acidic and, in part, sodic porphyries; the whole giving a strong impression of similarity to the pyritic deposits of Shasta County, California. On the other hand, there is no evidence of pneumatolytic deposition. On the contrary, the processes of replacement probably proceeded at moderate temperature and at moderate depth. Magnetite and pyrrhotite are present in only a few deposits, such as those at Cala, and opinions differ as to whether these minerals are due to later dynamometamorphic or contact-metamorphic processes or to original deposition.

The Pyritic Deposit of Rammelsberg, Germany[1]

The Rammelsberg deposit (Fig. 224), lies on the northern slope of the Harz Mountains, in Germany, near the town of Goslar. As is well known, it has been worked for copper ores since ancient times, the first records dating back to the tenth century. Its geological structure has been investigated by a number of authors, but its complete and detailed description is as yet a problem of the future. The most diverse explanations have been offered as to its mode of origin. By some, perhaps by a majority, including such well-known geologists as Bergeat and Klockmann, it has been considered as sedimentary deposit contemporaneous with the surrounding sedimentary rocks. Others, like Vogt, following Freiesleben and Lossen, explain it as a deposit from solutions immediately derived from igneous magmas. Still others, like Beck, are non-committal.

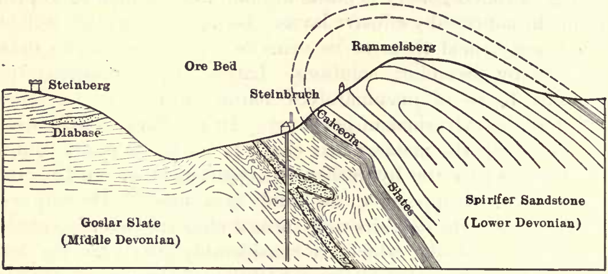

FIG. 224.—Cross-section of Rammelsberg, showing overturned anticline and supposed conformable position of the ore deposit. *After F. Klockmann.*

Geology and Structural Features.—The deposit is enclosed, apparently conformably, in Devonian rocks, which at Goslar appear as an overturned anticlinal and dip toward the north.

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[1] F. Klockmann, Berg- und Hüttenwesen des Oberharzes, 1895, p. 57.

J. H. L. Vogt, Über die genesis der Kieslagerstätten vom Typus Röros-Rammelsberg, *Zeitschr. prakt. Geol.*, 1894, p. 173.

W. Lindgren and J. D Irving, The origin of the Rammelsberg ore deposit, *Econ. Geol.*, vol. 6, 1911, pp. 303–313.

It lies in the so-called Goslar slates of the Middle Devonian; these slates are overlain by a thick series of Lower Devonian Spirifer sandstone, which makes up the summit of Rammelsberg Mountain, at the foot of which the mine is located.

The slates have suffered considerable deformation and the ore-body apparently follows their contortions more or less closely. The underground developments extend over a horizontal distance of about 2,000 meters and have attained a vertical depth of 380 meters from the level of the Richtschacht. It will be seen from this that mining has not yet penetrated to great depths, in spite of the fact that the deposit has been worked for nearly 1,000 years. The ore-body is divided horizontally into two parts, referred to as the old and the new beds; they are connected by a narrow and contorted seam, showing, however, beyond doubt that the two are really parts of one deposit.

The thickness of the "ore bed," as it is generally referred to, varies considerably; in places it swells to dimensions of as much as 30 meters, but this is rather due to folding and local enlargement. In most places the thickness is not over 2 or 3 meters and often only 0.5 to 1 meter. The dip is uniformly 45° to the southeast.

It is stated, even in the modern descriptions of the Rammelsberg, that the bedding or schistosity conforms exactly with the outline of the ore and with its banded structure. While true in places, this is certainly not a general characteristic of the deposit, which in part is absolutely unconformable to the stratification of the slate.

The ore banding everywhere follows with great faithfulness the outlines of the sulphide mass, whether these are smooth or irregular. For much of the distance above the third level the edge of the new ore-body and consequently the banding of the ore also are indeed parallel to the lamination of the enclosing rock. On the third level, however, the ore mass flattens out and crosses the lamination at a small angle, and again turns down parallel to it after intersecting the laminæ for a very considerable distance. In this portion of the ore-body the banding of the ore follows the edge of the sulphide mass and therefore makes the same angle with the lamination of the slates as the outline of the ore mass itself.

The Ores.—The principal minerals are zinc blende, chalcopyrite, galena, pyrite, and arsenopyrite, which are abundant approximately in the order enumerated. The gangue is almost entirely barite, but it rarely occurs in large quantities and often is entirely inconspicuous. Masses and veinlets of calcite are present in the surrounding slate, but rarely contain ore. On the whole the limits of the "ore bed" are sharply defined and the ores themselves are entirely or predominantly composed of sulphides. Alteration of the enclosing slates is rarely observed. At most there is a slight impregnation of pyrite.

The texture and composition of the ore vary with the locality, so that in one part copper will predominate, while in other places the ore carries mostly zinc blende with a little galena.

FIG. 225.—Nodules of barite and calcite (*A*) in banded sulphide ore from Rammelsberg, Germany. *After Lindgren and Irving.*

The intergrowths of these minerals are fine grained and intimate, which adds to the metallurgical difficulties of treating the ore. The most common texture is that of the banded ores consisting of dominant zinc blende with narrow and gently curved streaks of chalcopyrite and galena. In places the ore contains rounded nodules, generally of pyrite, around which the fine-grained streaks of zinc blende and chalcopyrite bend in regular curves. Pyrite shows a strong resistance to such deformation. Not uncommonly one finds rounded nodules, consisting of zinc blende and barite in granular form, coarser than that of the ordinary ore. Fig. 225 shows how the streaks of other sulphides surround and envelop these nodules. The pyrite nodules have often been noted, but without satisfactory explanation. Some observers have held them for fossil remains, in which the pyrite has replaced the shell of the organism. A goniatite has actually been found in the slates which form the easterly continuation of the ore-body, and, according to K. Andree,[1] the interior of the shell contained pyrite, quartz, barite, and calcite.

Origin.—The structural relations of the ore-body indicate that the deposit is a bedded vein—that is, a fissure vein lying in part, at least, conformable to the surrounding slates. The distribution and structure of the ore itself are inconsistent with the theory of sedimentary deposition. The structure is unique in ore deposits, but as to its interpretation there can be no reasonable doubt. The sulphides do not occur with their primary texture. The structure is that of a dynamometamorphic rock, in which all the constituents, except pyrite, have been drawn out into streaks which are intricately mingled. The appearance shown in the figure could be easily duplicated from any area of fine-grained schist resulting by pressure from an original granular rock. The different constituents have recrystallized and flowed under pressure.

At first glance it seems strange that the pyrite has acted so differently from the other constituents. The explanation of this behavior is found in some interesting experiments recently undertaken by F. D. Adams,[2] who shows that rock flowage is a function of hardness, so that the harder minerals are the less deformed. He finds that the limit of easily produced flowage lies in the vicinity of 5 or 6 in the scale of hardness. He crushed pyrite under a load of 43,000 pounds without its showing any trace of deformation by recrystallization. Minerals of lower hardness presented decided evidence of flow. The pyrite nodules in the Rammelsberg ores are simply residual parts of the original granular deposit which have been less deformed than the other sulphides.

The Rammelsberg deposit is then probably of epigenetic origin, but the structure of the ore has been profoundly changed by dynamometamorphism. While the surrounding slates are soft they evidently behaved quite differently from the sulphide mass.

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[1] *Zeitschr. prakt. Geol.*, 1908, p. 166.

[2] F. D. Adams, An experimental investigation into the action of differential pressure on certain minerals and rocks, *Jour. Geol.*, vol. 18, 1910, pp. 480–535.

The association of minerals with barite as the predominating constituent of the gangue tends to show that the deposit is not of the deep-seated type, but was formed at a medium depth below the original surface, probably within a few thousand feet of it.

The deposit may have been formed by ascending solutions derived from the neighboring batholith of granite, which is only 3 kilometers distant from the mine.

Similiar phenomena of flowage have been described recently from the Slocan district.[1]

CADMIUM ORES[2]

Almost the only cadmium mineral is the yellow grenockite (CdS), which is fairly common in the Joplin region, Missouri, as a yellow coating on zinc blende and disseminated in ozidized zinc ores coloring them yellow. Cadmium, probably as sulphide, is almost universally present in zinc blende. In the Joplin region the metal is contained to the extent of a trace up to 1 per cent., averaging 0.3 per cent., but many western ores also contain this metal in noteworthy amounts. It is found in high-temperature deposits, for instance, in those of Bodenmais, Bavaria, and also, as just stated, in zinc ores in limestone without igneous affiliations. The zinc deposits of Santander, Spain (p. 450), contain 0.4 per cent. Cd or more. Traces of cadmium have been found in some coals[3] and in mine waters. The allied metals, indium, thallium and gallium generally, accompany cadmium in small amounts. Germanium occurs in the tin deposits of Bolivia (p. 672).

Cadmium is produced in the United States since 1907, about 100 tons being the output in 1917. In addition 25 tons of cadmium sulphide were manufactured. The metal being more volatile than zinc is derived from the dust of the bag houses of lead smelters that treat zinc-bearing lead ores. Much more could be produced, especially from the electrolytic zinc plants if the demand required it. The German output of 40 to 50 tons a year is a by-product in the distillation of zinc in Silesia, the ores of

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[1] W. L. Uglow, *Econ. Geol.*, vol. 12, 1917, pp. 643–662.

[2] E. Jensch, Das Cadmium, etc., Samml chem. u. chem. techn. Vorträge, vol. 3. 1908, pp. 201–232.

C. E. Siebenthal, Cadmium, *Mineral Resources*, U. S. Geol. Survey, pt. 1, 1917, pp. 49–53.

[3] C. Schnabel, Handbook of Metallurgy, London, 1907.

that district containing at most 0.3 per cent. Cd. The price of cadmium ranges from $1 to $1.50 per pound.

Cadmium is used in the manufacture of silver ware and for easily fusible alloys, as a substitute for tin, in dental amalgam, etc. Cadmium sulphide is a brilliant yellow pigment.

ARSENIC DEPOSITS

Arsenic is of widespread occurrence, in volcanic exhalations, in the sea water, in many spring waters and spring deposits, and in some products of organic life. Nevertheless we rarely find pure arsenical ores in such abundance as to be of economic importance. About the only plentiful ore mineral of arsenic is arsenopyrite ($FeAsS$), which is found in many veins usually associated with quartz and gold. More rare is löllingite ($FeAs_2$), smaltite ($CoAs_2$) and niccolite ($NiAs$) and other cobalt and nickel arsenides; they occur, for instance, in the silver veins of Cobalt, Ontario (p. 627).

The minerals of arsenic are found in practically all classes of sulphide deposits of igneous affiliations, but are most plentiful in veins of the intermediate and high-temperature type.

In a few deposits like that of Hastings, Ontario[1] and Brinton, Virginia,[2] arsenious oxide (As_2O_3) has been recovered by roasting of arsenopyrite, but the great bulk is obtained as As_2O_3 in the flue dust of smelters using mixed ores in which occur the above mentioned arsenical minerals or some of the numerous sulpharsenides as tennantite ($Cu_8As_2S_7$), pearcite (Ag_9AsS_6), enargite (Cu_3AsS_4) and proustite (Ag_3AsS_3). There are several sulpharsenides of lead but they are rare.

In 1917, 6,151 tons of arsenious oxide was produced from flue dust. The price was from 10 to 16 cents per pound and the use is largely for killing insect pests or fungi, etc.

FLUORITE DEPOSITS[3]

Fluorite ($CaFl_2$) is the only simple fluoride occurring in nature It is a persistent mineral (p. 67) occurring in almost all deposits and formed at all temperatures. It rarely occurs in great abun-

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[1] J. W. Wells, *Eleventh Rept.*, Ontario Bur. Mines, 1902, pp. 101–122.

[2] F. L. Hess, *Bull.* 470, U. S. Geol. Survey, 1911, pp. 205–122.

[3] B. F. Burchard, Fluorite, *Mineral Resources*, U. S. Geol. Survey, pt. 2, annual issues, particularly 1916, which contains a good description of the English fluorite deposits.

dance. Most of the fluorite deposits worked are veins in limestone with igneous affiliations and formed at shallow or intermediate depths; among the associated minerals galena, pyrite and zinc blende are most common.

Fluorite is chiefly mined in the United States, in Illinois and Kentucky[1] from wide veins in Carboniferous limestone, shale and sandstone. The deposits contain also some quartz, barite, pyrite and galena. They are believed to stand in genetic connection with dikes of peridotite.

Fluorite is mined from veins in Colorado, in Boulder and Custer counties. The interesting hot spring deposit of Wagon Wheel Gap, in Mineral County, Colorado, is also of economic importance.

The English deposits in Derbyshire and Durham are likewise lead-bearing veins.

Fluorite is used as fluxing material, mainly in iron furnaces and for various industrial uses. Lenses are made from optically perfect material which is rare. It is also the raw material for hydrofluoric acid and other chemicals. In 1916, 155,735 short tons of fluorite were mined in the United States, mainly in Illinois and Kentucky, and brought an average price of $5.34 per ton. Regarding cryolite see p. 774.

SIDERITE DEPOSITS

Siderite occurs as a gangue mineral in many veins (p. 595). When it predominates and the veins are wide, as in southern Westphalia,[2] the vein filling is used as an iron ore. No such deposits are mined in the United States.

Near deposits in limestone this rock is often replaced by siderite (p. 616). Extensive replacements, probably accompanying intrusions, may result in deposits of siderite of economic value. At Eisenerz,[3] in Styria, a celebrated deposit of this kind is worked. Crushed and folded Devonian limestone form the material for the replacement. The ore, which averages 39 per cent. Fe, is worked in vast open cuts. The annual output is nearly 2,000,000 tons.

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[1] E. O. Ulrich and W. S. T. Smith, *Prof. Paper* 36, U. S. Geol. Survey, 1905.

J. F. Fohs, *Bull.* 9, Kentucky Geol. Survey, 1907; Econ. Geol., vol. 5, 1910, p. 377.

[2] Beyschlag, Krusch and Vogt, Ore deposits, Translated by Truscott, vol. 2, 1916, pp. 786-805.

[3] *Idem*, pp. 817-820.

CHAPTER XXVI

VEINS AND REPLACEMENT DEPOSITS FORMED AT HIGH TEMPERATURE AND PRESSURE AND IN GENETIC CONNECTION WITH INTRUSIVE ROCKS

GENERAL FEATURES

High-Temperature Minerals.—In the ore deposits described in previous chapters such minerals as the pyroxenes and amphiboles, the garnets, apatite, ilmenite, magnetite, pyrrhotite, tourmaline, topaz, the brown and green micas, the spinels, and the soda-lime feldspars are almost entirely absent. In the veins and replacement deposits formed at high temperature one or more of these minerals are commonly present, besides many other persistent ore and gangue minerals which are formed under widely varying conditions. In general, simple sulphides and arsenides prevail and are in many deposits associated with oxides, such as magnetite, ilmenite, and cassiterite. Some of the minerals enumerated above do not readily crystallize except with the aid of mineralizers (p. 760) for instance, certain compounds of boron, chlorine, or fluorine, which effect crystallization without always entering into the final compound.

In the presence of such mineralizers crystallization may take place at a much lower temperature than in dry fusion. In a magma high pressure is necessary to hold these substances in the fluid melt, which then is really a magmatic solution. The conceptions of solvent and solute are inapplicable, the various constituents of the magma being dissolved in one another. Under diminishing pressure, as during the ascent of magmas to higher levels, water and other mineralizers separate from the magmatic solution and carry with them certain constituents of the magma such as silica, some heavy metals, and alkaline metals. This "magmatic extract" may be in a state of aqueo-igneous fusion; or when the temperature is lowered the crystallization of some constituents may convert it into a fluid. In general, it is a hot mixture of fluid material and dissolved gases; many substances are doubtless above their critical points and would, if isolated, be in the non-compressible state known as a "perfect gas." This mixture would keep fluid far below the ordinary melting points of the minerals formed. The pegmatites, with their wealth of rare minerals, are considered as the product of consolidation of such aqueo-igneous melts and they also contain many of the minerals of the list given above.

Any magma always contains more or less of volatile matter which is not separated from it until the act of crystallization is in progress. Such material would ascend if suitable avenues of escape were provided and would probably mix with the earlier emanations and with water of surface origin.

This chemical evidence is supported by field evidence of the strongest kind. Practically all these deposits occur in or near bodies of intrusive rocks and have been exposed by deep erosion. They were, therefore, certainly formed at considerable depths below the surface. For some of them, like the contact-metamorphic deposits, cogent proof of the origin of the metals in the adjacent magmas can be given. On the other hand, the high-temperature veins at many places imperceptibly grade into those in which the magmatic origin is far less clear, thus giving in such places an almost complete line of transition from the rocks congealed from the magma, such as the pegmatite dikes, to the metal-bearing veins of the ordinary type.

In a given district these phenomena—the pegmatitic dikes, contact-metamorphic deposits, deep-seated veins, replacement deposits, and veins of the common type—all developed very soon after the intrusive activity and during a rather short and sharply defined epoch of metallization.

Some of the minerals enumerated on the previous page are dependent upon temperature only; such are the pyroxenes, spinel and magnetite; others like tourmaline, topaz, chondrodite, the micas, etc., contain a volatile constituent and require pressure and the presence of mineralizers for their formation. Some high-temperature deposits may, therefore, be formed comparatively near the surface and even in lava flows like certain rare tin deposits in rhyolite. Similar results may follow in case of intrusions near the surface when the temperature of the solutions are raised to such a degree that the vapor tension overcomes the pressure and fumaroles and "soffioni" result. In general, however, it will be found that most deposits with a mineral association indicating high-temperature have been formed at considerable or great depths.

Fumarolic action might result if, for instance, at a depth of 1,000 meters and consequent hydrostatic pressure of 100 atmospheres there existed a solution temperature of 330° C. The vapor pressure at this temperature would exceed the hydrostatic pressure. It is probable, however, that in such a case there could be no delicate banding of alternating minerals as so frequently characterizes the veins of shallow depth deposited by ascending hot waters.

The term pneumatolytic has often been used to indicate deposition above the critical temperature but it is also employed for any mineral formation by gases. It is probably best to discard a term of such an indefinite meaning.

When a mineral deposit only carries persistent minerals, like quartz and pyrite, there is no mineralogical criterion present to decide whether it belongs to the high temperature class. In many cases, though not always, mode of crystallization and geological criteria may solve the problem.

Metasomatic Processes.—The minerals enumerated in the beginning of this chapter generally appear in the metasomatically altered country rock and, to a less extent, in the fillings of the open cavities. The metasomatic action is often intense and leads to the development of coarse-grained aggregates. Sericitization still persists in some of these deposits though the foils of white mica may be larger and usually are associated with biotite, tourmaline and similar minerals. The total changes in feldspathic rocks are, however, often less pronounced than in the veins produced at lower temperature.

The carbonate rocks are always peculiarly susceptible to metasomatic processes and usually absorb large quantities of material from adjacent intrusives. Silicates rich in iron, like epidote, andradite (garnet), hedenbergite (pyroxene), cummingtonite (amphibole), and certain varieties of biotite, are frequently found in these deposits.

Under certain rarely occurring conditions andalusite develops in altered igneous rocks. B. S. Butler[1] describes such a case in the Beaver lake mining district, Utah, where a latite has been converted to quartz, muscovite and andalusite.

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[1] *Prof. Paper* 80, U. S. Geol. Survey, 1913, pp. 78–81.

Temperature and Pressure.—The actual temperature during deposition was probably rarely above 575° C., the inversion point of crystallization of quartz. It has been shown by O. Mügge, F. E. Wright, and E. S. Larsen[1] that the pegmatite dikes solidified at about this temperature. Where the rarer minerals have formed, the temperature, as indicated by the crystallographic behavior of the accompanying quartz, was commonly below that inversion point. The quartz veins and other deposits that by their mineral content show a relationship to the pegmatites are, as a rule, later than these and present many features which suggest that the process of cooling was further advanced. So, in a rough way, the temperature of deposition of this class must have been lower than 575° C., but in all probability higher than 300° C.

In the formation of the contact-metamorphic deposits, which were developed almost immediately after the actual intrusion of the magma, the temperature at the immediate contact may have been considerably above 575° C., attaining in acidic magmas 800° or 900° and in basic magmas 1,200° or 1,400° C.

It has been assumed that the heat necessary for the development of these deposits is derived from adjacent bodies of igneous rocks. The possibility cannot be denied that the same effect may be produced simply by the natural increment in temperature due to increase in depth. If a surface temperature of +25° C. and an increment of 1° C. for every 30 meters[2] are assumed, a depth of 10,200 meters, or about 33,600 feet, will be required for a temperature of 365° C., the critical temperature of water. Van Hise has shown that down to this depth even the hydrostatic pressure is sufficient to hold the water in the form of a liquid. Such observations as have been made in the Cordilleran region show that contact-metamorphic and other deposits of the type here called deep-seated have been formed much nearer to the surface, some of them at depths of 3,000 or 4,000 feet, the criterion being a rough measurement of the amount of erosion on the basis of known thickness of strata. It may be true for some problematical deposits of the Archean (for instance, the zinc deposit in limestone of Franklin Furnace, in New Jersey) that the rocks have been buried to a depth approximating 10,000 meters and

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[1] F. E. Wright and E. S. Larsen, Quartz as a geologic thermometer. *Am. Jour. Sci.*, 4th ser., vol. 27, 1909, pp. 421–427.

[2] C. R. Van Hise, *Mon.* 47, U. S. Geol. Survey, 1904, p. 567.

that, at that depth, they have been exposed to the metasomatic influence of magmatic gases, while they were at a considerable distance from igneous intrusions. Such deposits would be connecting links between igneous and regional metamorphism, and such a condition would explain the occasional occurrence of deposits of the contact-metamorphic type at a distance from known igneous bodies. The copper deposits of Ducktown, Tennessee, may furnish an example of this mode of formation, for here, although the ores are clearly of the contact-metamorphic type, there are no adequate igneous masses in the immediate vicinity which could have produced the metamorphism. The Ducktown district is one of intense regional metamorphism, and it is possible that magmatic gases of distant origin could have searched out the limestone beds and transformed the calcareous rock into ore.

As to the pressures actually existing our knowledge is slight. The hydrostatic pressure calculated by Van Hise would have little applicability, for at a relatively short distance below the surface the paths of underground water are probably effectively closed, and even where they are open the friction would be a factor of no mean importance. The pressure, therefore, at any considerable depth is probably far higher than that calculated from the weight of the water column. At a depth of 3,000 meters the hydrostatic pressure would be 300 atmospheres. Under purely static conditions the greatest pressure at any given point would be that indicated by the weight of the overlying rock column, or, for the depth just mentioned, equal to 810 atmospheres. Arching of resistant rocks might make this figure smaller; on the other hand, if the conditions are those of lateral stress it is possible that the actual pressure might be considerably higher and would then be measured by the strength of the buttress against which the pressure was applied.

If magmas and their differentiated gases invade the crust their pressure would be hydrostatic and could not exceed that of the static pressure of the overlying rock column without rupture of the rock. A contact-metamorphic deposit developing at a depth of 1,000 meters under a covering of limestones could, therefore, have been formed at a pressure of not more than 271 atmospheres.

Classes of Deposits.—The veins and replacement deposits formed at high temperature may be divided as follows:

A. Veins; replacement deposits not adjacent to intrusive contacts:

1. Cassiterite, wolframite, and molybdenite veins.
2. Gold-bearing veins and replacements.
3. Copper-tourmaline deposits.
4. Lead-tourmaline deposits.

B. Contact-metamorphic deposits (Chapter XXVII).

In these deposits we note again the remarkable connection of certain metals with certain igneous rocks. The tin, tungsten, and molybdenum veins, for instance, almost always appear in or near intrusions of acidic granites and porphyries.

The veins and replacement deposits carrying gold, copper, and iron are mainly connected with intrusive rocks of monzonitic or granodioritic character. In general, deposits of gold, copper, iron, tin, tungsten and arsenic are much more common than those of silver, lead, zinc and antimony.

Mode of Fissuring and Filling.—The question whether open spaces exist in the high-temperature veins has been discussed extensively. Under certain conditions at least it would seem improbable that open spaces could have existed where we now find cassiterite veins or gold-bearing quartz veins, for instance. Many investigators, such as W. O. Crosby, E. J. Dunn, F. B. Laney and Stephen Taber (p. 146) hold that the action of crystallization of minerals has forced the walls apart and thus provided space for the reception of ores. But aside from the problematic intensity of this force, such crystallization could hardly have produced perfect crystals or drusy structure. L. C. Graton,[1] in his description of the gold-quartz veins of the southern Appalachians, has suggested that the vein-forming solutions, representing the final products of emanation of a granitic magma, were injected under heavy pressure into the surrounding rocks along lines of weakness and so, like pegmatite dikes, made a space for themselves by opening their own fissures. The crystallization would be effected not so much by reduction of temperature and pressure, but rather by the disturbance of a nicely adjusted equilibrium of solubility and concentration by accession of substance dissolved from the wall rocks. This reasoning, which has much to commend it, would not be applicable where earlier fissures had established connection with the surface.

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[1] *Bull.* 293, U. S. Geol. Survey, 1906, p. 59

The texture of the veins is generally coarse grained and irregular. There may be a rude banding by deposition but nothing to equal the delicate concentric banding of the veins formed near the surface. Compare, for instance, Fig. 232 to Figs. 160 and 172. The structure of the composite veins or lodes is much like that of the deposits formed at intermediate depth. Compare Figs. 40 and 230.

THE CASSITERITE VEINS[1]

Mineral Association

The cassiterite veins form a rather sharply defined group, connected by transitions on the one hand with the copper-tourmaline veins and on the other hand with the wolframite and molybdenite veins. They present the constant association of such ore minerals as cassiterite, molybdenite, arsenopyrite, wolframite (also scheelite), bismuth, and bismuthinite, with less abundant pyrite, pyrrhotite, chalcopyrite, galena, and zinc blende. Among the gangue minerals quartz always predominates and is accompanied by lithium mica, fluorite, topaz, tourmaline, axinite, and apatite; more rarely beryl. Specularite, magnetite, and ilmenite are sometimes present. Of the primary carbonates, siderite is the only one which is reported from the cassiterite veins. On the other hand, the pyroxenes and amphiboles, as well as magnesium micas and garnets, are absent. Orthoclase is reported from several localities but does not assume the form of adularia. Chlorite is occasionally present. Kaolin and allied hydrous aluminum silicates are often recorded, but are probably products of secondary alteration near the surface, as are various hydrous arsenates and phosphates.

Cassiterite, the oxide of tin, is the principal ore mineral. The only other mineral containing tin which is of economic importance is stannite (Cu_2FeSnS_4), which is seldom found in pegmatites and in the cassiterite veins proper, but is an important ore in certain Bolivian veins.

Small quantities of tin, probably as cassiterite, are sometimes

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[1] H. G. Ferguson and A. M. Bateman, Geologic features of tin deposits, *Econ. Geol.*, vol. 7, 1912, pp. 209–262.

S. Fawns, Tin deposits of the world, London, 1907.

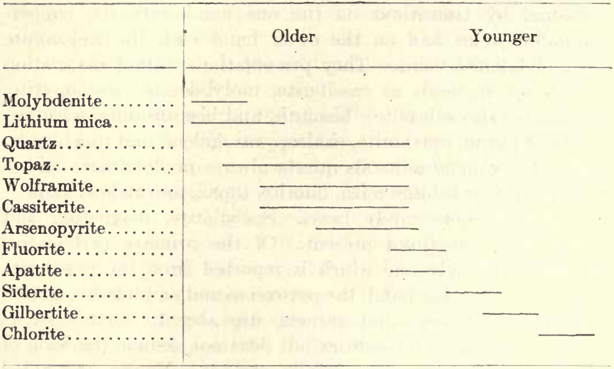
J. T. Singewald, Jr., Some genetic relations of tin deposits, *Econ. Geol.*, vol. 7, 1912, pp. 263–279.

contained in pyrite or zinc blende of other classes of veins—for instance, in those of Freiberg.

Cassiterite is extremely resistant to weathering, as shown, for example, by its occurrence in placers. It is held by some authors that the so-called fibrous tin ore or "wood tin" which is often found in placers is a product of alteration of cassiterite, but the question does not seem to be definitely settled. If secondary, it is probably derived from stannite.

The paragenesis typical of many cassiterite veins is indicated in the following table.

SUCCESSION OF MINERALS IN THE CASSITERITE VEINS OF SAXONY
(After R. Beck)



The tin-bearing veins appear in or near granites (though not all granites contain them), or in their acidic porphyries. Only exceptionally, as in Mexico, are they connected with rhyolitic rocks. The tenor of the ores is usually low, in some ores as low as one-half of 1 per cent. of tin.

Some cassiterite veins contain bismuth and tungsten minerals in commercial quantities, and considerable copper is often present. A little silver and a trace of gold are found even in the vein of Cornwall, while in the Bolivian veins silver minerals occur in important amounts.

In 1916 the world's output of tin was 136,000 short tons most of which came from the Malay Peninsula and adjacent regions. Bolivia yielded 23,500 tons. Only 140 tons were produced from domestic ores in the United States; this came mainly from Alaska.

Metasomatic Processes[1]

General Features.—The tin ores generally appear in distinct fissure veins or composite lodes; in part they fill open cavities and in such ores a banded structure may be emphasized by the deposition of lithium mica in coarse foils along the walls. Often, however, the fissures are merely narrow breaks and the ore is chiefly disseminated in the adjoining altered country rock. In ores of this kind also a rude banding may result from the accumulation of tourmaline or cassiterite along certain lines parallel to the fissure.

The metasomatic alteration is strong and characteristic, resulting in coarse-grained rocks which contain muscovite, quartz, and topaz or tourmaline and to which the name greisen is usually applied (Fig. 226). Where tourmaline takes the place of topaz we may speak of tourmaline greisen or luxullianite (Cornwall). Granite, granite porphyry, clay slate, calcareous shale, limestone, and diabase are affected by this mode of alteration where they form the country rock of the veins, but the development differs

FIG. 226.—Thin section of greisen from Banka, derived from granite. *g*, Lithium mica; *q*, quartz; *z*, cassiterite; *t*, topaz; stippled spots in mica consist of zircon and rutile, surrounded by pleochroic rings. Magnified 45 diameters. *After R. Beck.*

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[1] Besides the special papers, see F. Zirkel, Lehrbuch der Petrographie, vol. 2, 1894, pp. 118–127.

in the different rocks. Complete silicification of the wall rocks is a phase of subordinate importance.

While the total changes in composition of the original rock may be much less pronounced than in other veins, the metasomatic process is evidently more intense, pointing to a greater degree of heat and especially energetic action of the mineralizers.

In granites and porphyries adjoining the more common or Saxon type of veins the feldspars and the brown mica are replaced by quartz, topaz, and muscovite, in large crystals or foils. Chlorite is sometimes present. Topaz may also replace quartz grains. Sometimes crystals or radial aggregates form. Cassiterite, wolframite, and more rarely sulphides appear as accessories in the greisen, which spreads out from the fissure plane for a few inches or a few feet.

The quartz porphyry dikes of Mount Bischoff, in Tasmania, contain cassiterite, with much topaz and subordinate tourmaline. The groundmass is changed to a topaz-quartz aggregate, while the feldspars are transformed to cassiterite, pyrite, pyrrhotite, arsenopyrite, and fluorite. The quartz phenocrysts remain intact. In the final product quartz and minute crystals of zircon are the only constituents which have withstood the altering processes. Siderite appears in places as a metasomatic product.

Metasomatic Processes in the Deposits of Cornwall.—The Cornwall granites,[1] which consist mainly of quartz, acidic feldspars muscovite, and biotite, also carry some tourmaline, topaz, and fluorite, which Flett considers of magmatic origin. The greisen along the Cornwall veins consists chiefly of granular quartz and muscovite, often with aggregates of topaz spreading through the partly altered feldspars. Fluorite is occasionally present. The albite is more resistant than orthoclase or perthite. Some of the secondary quartz is filled with liquid inclusions containing small cubical crystals and mobile bubbles. Excellent pseudomorphs of cassiterite after orthoclase have been found at Cornwall; pseudomorphs of tourmaline after feldspar and of topaz and cassiterite after quartz are also described by Flett.

The elvans, or quartz porphyries, are also altered to quartz, tourmaline, topaz, and fluorite. Kaolin where present appears to be due to a later process.

The greisen is essentially a vein formation in Cornwall and is not known to occur in broad masses or in patches either within

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[1] J. S. Flett, Explan. Sheet 374, Geol. Survey England

the granite or along the contact. The occurrence along the contact is typical of the tourmaline rock, which, however, in places also appears along the veins. Tourmalinization is frequently superimposed upon the normal contact-metamorphic rocks, of which hornfels is the most common. There is not much evidence to show the relative age of the two processes, but the greisen was probably developed later than the tourmaline rocks. Both are held to have been formed before the interior of the granite was fully crystallized.

Considerable portions of the granites of Cornwall have been altered by kaolinization, but Flett believes, with good reason, that this process took place at a lower temperature than the development of topaz, tourmaline, and white mica. Kaolinized rocks appear mainly in the central parts of the granite masses, while the tourmaline rock, usually called "schorl," and greisen lie mostly along the peripheral parts. The kaolinized portions often form pipes having rudely circular outlines; the granite is altered to kaolin, muscovite, and quartz, and the product rarely contains cassiterite. The composition differs little from that of the granites. It seems probable that these kaolin deposits are due to the oxidation of distinctly later pyritic impregnations, by the action of the liberated sulphuric acid on the feldspar and sericite. It should also be recalled that topaz is rather easily changed to products allied to kaolin.

Development of Greisen.—In the following table three sets of analyses are given showing the development of greisen in Saxony, in New South Wales, and in Cornwall. On the whole, the processes are identical. If, as seems probable, there has been but little change in the total mass of the alumina, the analyses are roughly comparable in their present form. Silica has also remained fairly constant. The additions consist of iron, lithium, tin, fluorine, and boron, the iron evidently entering in a silicate. Calcium, of which but little is present, is strongly leached in the rocks represented by analyses IV and VI; the evidence as to magnesia is less conclusive. Sodium and potassium have both been abstracted, the former more than the latter, but neither is wholly removed. In this feature the metasomatic alteration of tin veins differs from that along veins formed at lower temperature, in which we often find a strong concentration of potassium and an almost complete removal of sodium. The specific gravities of the rocks are not determined.

ANALYSES OF GRANITES AND GREISENS DEVELOPED FROM THEM

	I	II	III	IV	V	VI
SiO_2	74.68	70.41	70.17	69.42	76.69	78.47
TiO_2	0.71	0.49	0.41	Trace		0.22
SnO_2	0.09	[a]0.49				0.08
Al_2O_3	12.73	14.86	15.07	15.65	10.89	11.50
Fe_2O_3		1.42	0.88	1.25	0.76	2.64
FeO	3.00	5.09	1.79	3.30	0.39	1.05
CuO	0.50					
MnO		0.29	0.12	0.39		
CaO	0.09	0.21	1.13	0.63	1.73	Trace
MgO	0.35	0.09	1.11	1.02	0.18	0.49
K_2O	4.64	3.01	5.73	4.06	2.97	1.17
Na_2O	[b]1.54	0.98	2.69	0.27	5.35	1.99
Li_2O			0.11	0.81		
H_2O+	1.17		0.70	0.54	0.13	0.23
H_2O-			0.18	0.06	0.37	1.17
P_2O_5			0.34	0.40		
MoS_2						0.80
Cl			0.06	Trace		
F		3.10	0.15	3.36		Present
S			0.04			
B_2O_3			Strong trace	0.59		
			100.68	101.75		
Less O for F and Cl			0.07	1.41		
Total	99.50	100.44	100.61	100.34	99.46	99.81

[a] As cassiterite 0.43; in mica, chemically combined, 0.06.

[b] Including lithia.

I. Fresh granite, Altenberg, Saxony. K. Dalmer, Explanations to the section Altenberg-Zinnwald. Geol. map Saxony.

II. Greisen, Altenberg, Saxony. K. Dalmer, *idem.*

III. Fresh Lamorna granite, Lands End, Cornwall. W. Pollard, analyst.

IV. Greisen, with tourmaline and topaz. St. Michaels Mount, Lands End, Cornwall. W. Pollard, analyst. Explan. Sheets 351 and 358, Geol. Survey England.

V. Fresh "acidic granite," New England, New South Wales. L. A. Cotton, analyst. *Proc.*, Linnean Soc. N. S. W., 34, pt. 2, 1909, pp. 220–238.

VI. Greisen, cassiterite vein near Inverell, same locality. L. A. Cotton, analyst. *Idem.*

The composition of a normal greisen from the Erzgebirge, in Saxony, is, according to Dalmer, as follows: Quartz, 50.28; topaz, 12.14; lithium mica, 36.80; and cassiterite, 0.43 per cent.

Alteration of Sedimentary Rocks.—The alteration of sedimentary rocks proceeds somewhat differently. In Cornwall the argillaceous slates are tourmalinized, forming "cornubianite" (Fig. 227), the biotite and muscovite being replaced by that mineral. Andalusite and cordierite also disappear, but rutile, ilmenite, and magnetite remain. The result is an aggregate of quartz and tourmaline, which well preserves the original structure of the rocks. In place tourmaline-albite rocks are formed.

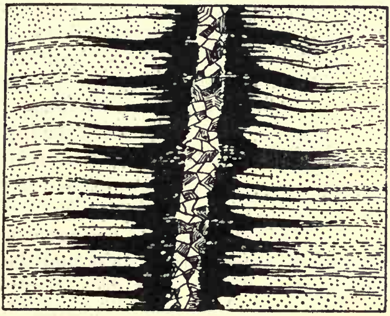

Fig. 227.—Vein of quartz, cassiterite, and tourmaline traversing Paleozoic slates, in which the argillaceous bands are replaced by tourmaline; the siliceous bands in the slate are not altered. Belowda Beacon, Cornwall. Natural size. *After D. A. MacAlister.*

The calcareous rocks, as well as the greenstones, yield mainly axinite in large brown crystals, also pyroxene, actinolite, epidote, garnet, zinc blende, pyrite, apatite, specularite, titanite, and tourmaline, but no topaz. At Mount Bischoff, Tasmania, the probably non-calcareous schists and slates are in part changed to tourmaline fels, containing also cassiterite, pyrite, arsenopyrite, pyrrhotite, fluorite, calcite, siderite, and pyrophyllite.

H. Tronquoy[1] describes tin-bearing veins from Villeder, Morbihan, France, which are accompanied by albitization of granite, without tourmaline or topaz while tourmaline develops in adjacent clay slate.

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[1] Comptu Rendu, 154, 1912, p. 899.

The metasomatic development in the sedimentary rocks is of great interest, for it connects in the closest manner the effects of the ore-bearing solutions with those in contact-metamorphic deposits. Regarding the cassiterite deposits of Alaska and Pitkäranda and their close association with contact-metamorphism, see p. 741.

Origin of Tin-Bearing Veins

The tin-bearing veins occupy a most important position as connecting links between the deposits of igneous and aqueous origin.

The occurrence of cassiterite as a primary constituent of granite is recalled, as well as its occasional appearance in the pegmatitic druses of granite—for instance, those in Elba described by G. vom Rath.[1] We may further recall the appearance of cassiterite in economic quantities in pegmatite dikes (p. 768) and its occasional occurrence in base-metal veins—for instance, in those of Freiberg and in the lower levels of the Przibram veins. The occurrence of tourmaline in cassiterite veins is also important in view of the transitions to the chalcopyrite-tourmaline type.

These facts were realized at an early date by A. Daubrée and E. de Beaumont.[2] To J. H. L. Vogt[3] belongs much credit for his resuscitation of these meritorious ideas and the addition of important investigations. The extensive literature regarding the occurrence of tin deposits has been summarized by F. L. Hess and L. C. Graton.[4]

Daubrée and de Beaumont argued from the close association of cassiterite veins and acidic granites that there must be some genetic connection between them and concluded that the veins were deposited by emanations from these magmas. This conclusion has been adopted and confirmed by almost all geologists who have studied these deposits.

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[1] G. vom Rath, *Zeitschr.* Deutsch. geol. Gesell., 1870, p. 646.

[2] A. Daubrée, Sur le gisement, etc., des amas de mineral d'étain, *Annales des Mines*, 3d ser., vol. 20, 1841, and other later papers.

E. de Beaumont, Notes sur les émanations volcaniques et métallifères, *Bull.* Soc. Min. France, 2d ser., vol. 4, 1847, p. 65.

[3] J. H. L. Vogt, Zinnstein-Ganggruppe, *Zeitschr. prakt. Geol.*, 1895, pp. 145–156, and other places.

[4] F. L. Hess and L. C. Graton, The occurrence and distribution of tin, *Bull.* 260, U. S. Geol. Survey, 1905.

F. L. Hess and Eva Hess, Bibliography of the geology and mineralogy of tin, *Smithsonian Misc. Coll.*, vol. 58, No. 2, 1912.

The general occurrence of cassiterite in well-defined veins shows clearly that the granite was consolidated when these fissures were broken, even if it is probable that the whole granitic intrusion had not congealed and that liquid magma still existed below the vein. The veins occur usually either in the center or along the contacts of the granitic masses, but some of them extend into the adjoining sedimentary rocks. Finally, it is clear that the development of these veins cannot be identified with the contact-metamorphism, for they are distinctly later and their metasomatic effects are superimposed upon the products of contact-metamorphism. Granitic magmas apparently do not easily part with their volatile constituents, which are mainly expelled during consolidation.

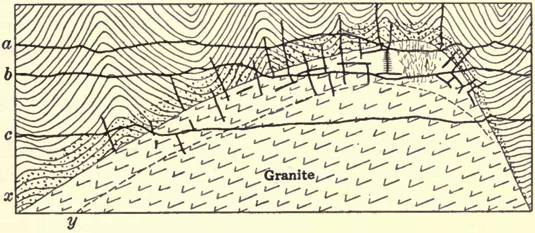

Fig. 228.—Ideal section of granite intrusion showing probable development of tin-bearing veins and the effect of successive erosion levels, *a*, *b*, and *c*; *x*, metamorphic aureole; *y*, inner limit of mineralization. *After Ferguson and Bateman.*

The general distribution of tin-bearing veins in relation to an intrusive mass of granite gradually eroded is shown in the diagram, Fig. 228.

The constant presence of boron and fluorine compounds, as well as those of phosphorus, in the tin veins is, of course, of the highest importance. During the short but intense epoch of metallization the temperature must have been high, probably above the critical temperature of water, and therefore the deposition took place under "pneumatolytic" conditions. In the absence of exact data as to the temperature, pressure and constitution of the vein-forming agents, the insistence upon pneumatolysis has, however, little value. Just how much water was present is problematical; certainly there was some, as shown by the hydroxyl radicle in muscovite and topaz, and by the presence of aqueous inclusions in quartz crystals.

Daubrée's reaction, inferred as probable from experiments with the chloride is as follows:

$$SnFl_4+2H_2O=SnO_4+4HFl.$$

The metasomatic development of topaz has been taken to indicate that free hydrofluoric acid was present in the solution, but this must be regarded as very improbable. We may better frankly state that at present we are in the dark as to the exact formulas of the reaction.

Both in Saxony[1] and in Cornwall lead-silver veins occur at some distance from the center of intrusions and the cassiterite veins, and also transitions between them. This would tend to show that these metals were less easily precipitated and were carried farther away from their sources than were the tin, tungsten, etc.

A similar relation exists in the Cordilleran region between the copper-bearing contact-metamorphic deposits and the lead-silver replacement deposits in limestone.

The Cassiterite Veins of Cornwall, England

Literature.—An extensive literature exists on the subject of the tin lodes of Cornwall, for they have been repeatedly studied by geologists since they were first described in 1839 by H. T. de la Bèche, and in 1843 by W. J. Henwood. Many articles by J. H. Collins and C. Le Neve Foster were published about 30 to 35 years ago. The deposits have been studied recently by the Geological Survey of England, and the results are published in a series of memoirs.[2]

The renowned mining region of Cornwall, known to the ancients for the treasures of tin which it contained, furnishes one of

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[1] K. Dalmer, *Zeitschr. prakt. Geol.*, 1894, p. 321.

[2] D. A. MacAlister, Geological aspects of the lodes of Cornwall, *Econ. Geol.*, vol. 3, 1908, pp. 363–380.

J. B. Hill, D. A. MacAlister, and J. S. Flett, Geology of Falmouth and Truro, *Mem.* Geol. Survey England, Explan. Sheet 352, 1906.

Ussher, Flett, et al., The geology of Bodmin and St. Austell, *Idem*, Explan. Sheet 347, 1909.

C. Reid and J. S. Flett, The geology of the Lands End district, *Idem*, Explan. Sheets 351 and 358, 1907.

C. Reid and J. S. Flett, The geology of Camborne and Redruth, *Idem.*

the most remarkable instances of the dependence of ore deposits on the distribution of igneous rocks. The folded Paleozoic slates and sandstones are intruded by four main granitic batholiths of moderate dimensions (Fig. 229), and the tin deposits cluster characteristically in the marginal zones of these granitic intrusions, both in slates and in granites. The slates, or killas, as they are locally called, in part overlie the granite, forming in places the roof of the batholith. The intrusion is post-Carboniferous and pre-Triassic in age, and the veins were formed shortly after the intrusive activity, probably while the rock still remained hot. Before the vein-forming epoch a series of dikes of granite porphyry (elvans) intersected granite and slate. The granite is a typical rock of its kind, containing, in order of crystallization, zircon, apatite, magnetite, biotite, muscovite, oligoclase, albite, perthite, and quartz. Tourmaline, topaz, and fluorite are occasional accessories in the crystallization of the magma.

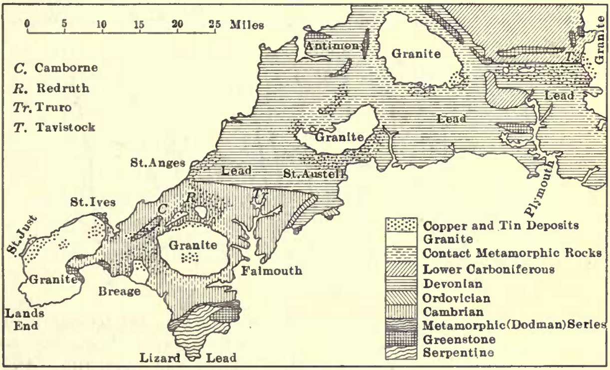

FIG. 229.—Geological map of the peninsula of Cornwall, England. *After D. A. MacAlister.*

The tin and copper veins are the older and were followed by a later though much less important series of veins, containing lead, silver, and sometimes also cobalt, nickel and uranium. A little gold is present, both in the lead-silver veins and in the cassiterite veins. Many of the veins are of complex structure and may properly be called lodes. Some of them are traceable for 2 or 3 miles, or even more. The width of infilled fissures and altered rocks averages 3.5 feet, but in places, especially in the slates, reaches 50 feet. The general direction of the veins is northeast. Stockworks of irregular veinlets also occur. Faulting has taken place along many of the fissures. The veins are simple or composite (Fig. 230); reopening of fissures was evidently a common phenomenon. The filling is mainly of quartz, frequently with comb structure and, in the upper parts of the lodes, also with drusy cavities. Chlorite, fluorite, scheelite, zinc blende, molybdenite and bismuthinite are also found in these veins. Magnetite and specularite are reported, but are rare.

The alteration spreading into the country rock from the fissures is characteristic and, when effected from a series of closely spaced fissures, may produce a banded appearance (capel); tourmalinization is the usual mode of alteration in the granite, resulting in an aggregate of tourmaline and quartz. In calcareous rocks or greenstones the altered zone along the veins contains axinite, pyroxene, garnet, and some tourmaline.

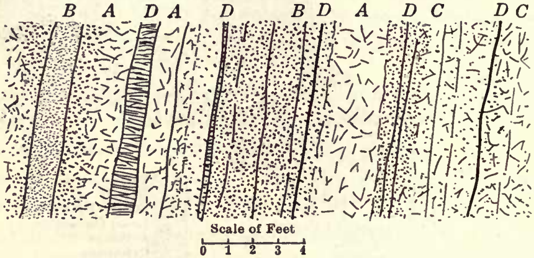

FIG. 230.—Diagrammatic section of the main lode at the 300-foot level, Bunny mine, St. Austell, Cornwall. *A*, Kaolinized granite; *B*, stanniferous greisen; *C*, silicified granite; *D*, veins of quartz and cassiterite. *After Ussher, Flett, et al.*

In the slates the lodes often contain much crushed and brecciated material; sometimes cassiterite and tourmaline following bedding planes impregnate the slates for some distance on both sides of the lode. On the whole the copper ores are confined to the lodes in the slates. Where the lodes break into the granite the tin ore makes its appearance (Fig. 231).

The main lode of the Dolcoath and Carn Brea contained copper ores down to a depth of 1,000 feet, mainly in the slates. Below this depth the tin ore predominated and is now worked at a depth of 3,000 feet. According to Hill and MacAlister the lode at the bottom of the mine is 42 feet wide and contains 1 to 3 per cent. of cassiterite. The lodes of Wheal Vor were of enormous value in the killas, but proved worthless in the granite. At the Great Work mine, not far distant, these relations were reversed. The great ore shoots of both mines plunge eastward approximately parallel to the subterranean contact surface of the granite.

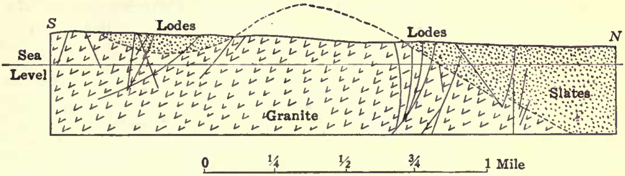

Fig. 231.—Section from Feltrick to New North Pool, Cornwall, showing relation of granite and slate and the lodes intersecting them. *After Hill and MacAlister.*

Cassiterite Veins of Saxony[1]

Saxony and the adjacent part of Bohemia contain several tin-producing districts, the ores occurring in or near granites of post-Carboniferous age. The deposits are now of little importance, but have an interesting place in the history of the study of ore deposits.

At Altenberg the ores occur in a stockwork, about 3,000 feet in diameter, of small veins cutting across the granite and the adjacent, older granite porphyry; the whole mass of rock is altered to a greisen, containing a little cassiterite and arsenopyrite. The characteristic minerals occurring in the veinlets are quartz, bismuth, bismuthinite, pyrite, chalcopyrite, molybdenite, zinc blende, wolframite, fluorite, tetrahedrite, magnetite, and specularite. The ore, according to Dalmer, continued only to a depth of about 700 feet.

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[1] The older literature includes the papers by B. v. Cotta, H. Müller, E. Reyer, and A. W. Stelzner.

R. Beck, Einige Beobachtungen im Gebiete der Altenberg-Zinnwalder Zinnerzlagerstätten, *Zeitschr. prakt. Geol.*, 1896, pp. 148–150.

K. Dalmer, Der Altenberg-Graupener Zinnerzlagerstättendistrikt. *Zeitschr. prakt. Geol.*, 1894, pp. 313–322; see also *idem*, 1895, p. 228; 1896, p. 1; 1897, p. 265; 1900, p. 297.

J. T. Singewald, Jr., The Erzgebirge tin deposits, *Econ. Geol.*, vol. 5, 1910, pp. 166–177, 265–272

At Zinnwald the veins are likewise in granite, which with flat contact breaks through quartz porphyry. The numerous fissures are approximately parallel to the contact, and the veins are formed mainly by filling, sometimes with excellent banded structure by the development of mica next to the wall; they contain a considerable amount of wolframite (hübnerite), besides the usual minerals accompanying the cassiterite (Fig. 232).

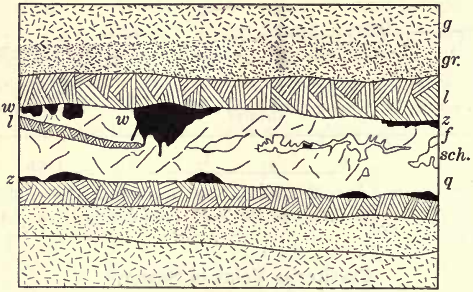

FIG. 232.—Diagrammatic section of vein at Zinnwald, Saxony. *g*, Granite; *gr*, greisen; *q*, quartz; *l*, lithium mica; *z*, cassiterite; *w*, wolframite; *f*, fluorite; *sch*, scheelite. *After R. Beck.*

Tin Deposits in Other Countries

Cassiterite veins have been found in the Transvaal, New South Wales, Queensland, Western Australia, Tasmania, Bolivia, Mexico, and Alaska. In the United States they are rare. A vein in granite has recently been worked near El Paso, Texas, on the east side of the Franklin Mountains.[1] Some low-grade veins occur in the Temescal Mountains, near Riverside, California.[2] The tin-bearing deposits of the Appalachian region and of the Black Hills of South Dakota are mainly pegmatite dikes, though a few quartz veins with cassiterite occur in the Appalachian region.[3] The occurrences in Alaska are described on p. 741.

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[1] W. H. Weed, The El Paso tin deposits, *Bull.* 178, U. S. Geol. Surv., 1901.
G. B. Richardson, Tin in the Franklin Mountains, *Bull.* 235, U. S. Geol. Survey, 1906, pp. 146–149.

[2] H. W. Fairbanks, The tin deposits of Temescal, *Am. Jour. Sci.*, 4th ser., vol. 4, 1897, pp. 39–42.

[3] L. C. Graton, Reconnaissance of some gold and tin deposits of the southern Appalachians, *Bull.* 293, U. S. Geol. Survey, 1906

The largest part of the tin production of the world is derived from the Malay Peninsula and from the islands of Banka and Billiton, off the coast of Sumatra, but the output from these localities comes mainly from placers. In the Malay States granite invades post-Triassic limestone and schist, and tin-bearing veins in these rocks have been described by Penrose.[1] On the islands granite is intruded in slate; both rocks contain tin veins which are noteworthy in that they carry magnetite.

Interesting tin deposits have been described as occurring in the Transvaal and New South Wales. They occur in granite and are roughly cylindrical "pipes" consisting of highly altered granite with disseminated cassiterite and tourmaline.[2] Veins and pipes of cassiterite accompanied by sulphides, specularite, tourmaline, and ankerite or siderite occur in the quartzite of the Rooiberg district[3] in the Transvaal. The deposits are several miles distant from the intrusive contact.

The principal tin deposits of Tasmania are those of Mount Bischoff, where schists are intruded by dikes of granite porphyry, both rocks being rich in metasomatic tourmaline and topaz. The whole forms a weathered mass of rock traversed by cassiterite veins—a stockwork, large portions of which averaged 2 to 3 per cent. tin. Magnetite, fluorite, pyrrhotite, zinc blende, wolframite, and siderite are mentioned as accompanying minerals.[4]

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[1] R. A. F. Penrose, The tin deposits of the Malay Peninsula, *Jour. Geology*, vol. 2, 1903, pp. 135–154.

W. Wolff, J. B. Scrivenors Arbeiten über die Geologie von britisch Malaya, *Zeitschr. prakt. Geol.*, 1911, pp. 152–158.

J. B. Scrivenor, The topaz-bearing rocks of Gunong-Bakau, *Quart. Jour.*, Geol. Soc. London, vol. 70, 1914, pp. 363–381.

W. R. Jones, The origin of topaz and cassiterite at Gunong-Bakau, *Geol. Mag.*, vol. 3, 1916, pp. 255–260.

R. D. M. Verbeek, Ueber die Zinnerzlagerstätten von Bangka und Billiton, *Zeitschr. prakt. Geol.*, 1899, pp. 134–136.

C. T. Groothoff, De primaire Tinertsafzettingen van Billiton, Dissertation, Delft, 1916.

[2] H. Kynaston and E. T. Mellor, The geology of the Waterbury tin field, *Mem.* 4, Transvaal Geol. Survey, 1909.

L. A. Cotton, The tin deposits of New England, New South Wales, *Proc.*, Linnean Soc. N. S. W., vol. 34, 1909, pp. 733–781.

[3] R. Recknagel, *Trans.* Geol. Soc. S. Africa, vol. 11, 1908, pp. 83–106.

[4] W. von Fircks, Die Zinnerzlagerstätten des Mount Bischoff, *Zeitschr.* Deutsch. geol. Gesell., Bd. 51, Heft 3, 1899, pp. 431–465.

P. Krusch, Die Zinnerzlagerstätten des Mount Bischoff, *Zeitschr. prakt. Geol.*, 1900, pp. 86–90.

The Bolivian veins,[1] which center in the mining districts of Oruro and Tres Cruces, appear in Devonian slates, intruded by dikes of granite porphyry. A. W. Stelzner, who first described them, drew attention to certain unusual features consisting in the association with silver minerals and the occurrence of much pyrite, but W. R. Rumbold and M. Armas have shown that the veins are similar to the normal tin veins and that the country rock is extensively tourmalinized. In some of the veins stannite ($FeCu_2SnS_4$) is present, as well as ruby silver, stephanite, tetrahedrite, zinc blende, arsenopyrite, bournonite, wolframite and siderite. Germanium minerals such as argyrodite ($3Ag_2S.GeS_2$) was found at Potosi, and canfieldite, an argyrodite containing tin, at La Paz. The tin ores of Bolivia are in part of high grade and yield a production of great importance.

The tin-bearing district of Zeehan, Tasmania,[2] contains deposits of various kinds which appear to show an unusually complete series of transitions. Silurian sediments are intruded by granite. There is a gradation from cassiterite veins, with tourmaline, in granite, to contact-metamorphic deposits, containing copper, lead, and zinc, in which cassiterite has been found, and from these to normal banded veins containing pyrite, chalcopyrite, galena, and stannite, and finally to galena-siderite veins. In other words, the gradation is one from high-temperature deposits to those formed in the cooler zone with a corresponding change of minerals deposited.

Minor veins of cassiterite are sometimes found in rhyolite flows, in which they were evidently formed shortly after the consolidation of the rock. We have here then high temperature deposits originating near the surface. It may be recalled that topaz is sometimes found in lithophysæ in rhyolite and these phenomena indicate a retention by the lavas of certain of their volatile constituents until consolidation of the rock. Such deposits contain concretionary cassiterite and "wood tin" associated with hematite, chalcedony, and opal, sometimes also with

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[1] A. W. Stelzner, Die Silber-Zinnerzlagerstätten Boliviens, *Zeitschr* Deutsch. geol. Gesell., vol. 49, 1897, pp. 51–142.

W. R. Rumbold, The origin of the Bolivian tin deposits, *Econ. Geol.*, vol. 4, 1909, pp. 321–364.

M. Armas, *Ann. des Mines*, 10th ser., vol. 20, 1911, pp. 149–213.

[2] W. H. Twelvetrees and L. K. Ward, *Bull.* 8, Dept of Mines, Tasmania, 1910.

wolframite and bismuth minerals. Occurrences of this kind have been described from Mexico[1] and from Nevada.[2]

WOLFRAMITE VEINS

Wolframite, including the tungstate of iron (ferberite) and the tungstate of manganese (hübnerite), has a field of occurrence similar to that of cassiterite. It appears in igneous rocks, in pegmatites, in cassiterite veins, and sometimes with quartz and bismuth minerals in veins which are evidently of the deep-seated type and allied to the tin veins. But, unlike cassiterite, wolframite also appears abundantly in veins formed under much more moderate temperature and pressure—for instance, in those of Boulder County, Colorado (p. 620). Small quantities of hübnerite are found in veins formed near the surface, as in those of Tonopah, Nevada, and Cripple Creek, Colorado. The principal production in the United States is derived from Boulder County, Colorado.

Wolframite veins of the deep-seated type have been described from the Deer Park district,[3] in Washington and from Sauce[4] in the Sierra de Cordova in Argentina. Wolframite lodes of great importance have lately been developed in the Tavoy district, Lower Burma,[5] where they occur in granite and metamorphic schist and contain in addition quartz, mica, tourmaline, columbite, arsenopyrite, pyrite, bismuthinite and galena. At the same place wolframite is also recovered from alluvial deposits. In 1916 the mines of Lower Burma produced about 3,000 tons of wolframite concentrates, which is about one-third of the annual output of the world. The crude ore is said to average 1.3 per cent. WO_3 per ton.

Regarding molybdenite veins see p. 777.

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[1] W. R. Ingalls, The tin deposits of Durango, Mexico, *Trans.*, Am. Inst Min. Eng., vol. 25, 1896, pp. 146–163; vol. 27, 1898.

E. Halse, *idem*, vol. 29, pp. 1900, 502–511.

E. Wittich, Zinnerze in der Sierra von Guanajuato, *Zeitschr. prakt. Geol.*, 1910, pp. 121–123.

[2] Adolph Knopf, Tin ore in northern Lander Co., Nevada, *Bull.* 640 U. S. Geol. Survey, 1916, pp. 125–138.

[3] Howland Bancroft, *Bull.* 430, U. S Geol. Survey, 1910, pp. 214–216

[4] Bodenbender, *Zeitschr. prakt. Geol.*, 1894, pp. 409–414.

[5] A. W. G. Bleeck, *Records* Geol. Survey India, vol. 43, pt 1, 1913.

E. M. Lefroy, *Min. Mag.*, vol. 25, 1916, pp. 83–100.

H. D. Griffiths, *idem*, vol. 26, 1917, pp. 60–65.

GOLD-QUARTZ VEINS

Gold-bearing veins of a deep-seated type are found in many regions in the pre-Cambrian and earliest Paleozoic rocks of the American continents. They appear in the gold belt of the Appalachian States, mainly from Maryland to Alabama; at various places in the Western States, particularly in South Dakota and New Mexico; in Ontario and Quebec; and finally in Brazil.

The Veins of the Southern Appalachians[1]

The placer deposits of the southern Appalachians have, since their discovery, about 1800, yielded gold estimated at $30,000,000. The veins from which the placers were derived proved less productive, though they have been profitably worked at many places in North and South Carolina, at Dahlonega and the Franklin mine in Georgia, and at the Hogback mine in Alabama. All the deposits are not of the deep-seated type; there are some which more closely approach the normal gold-quartz veins similar to those of California, but even in these deposits certain features indicate deposition at higher temperatures. Others, like those described by Taber from Virginia, seem to be related to pegmatite dikes. The veins are contained in crystalline rocks, usually more or less schistose, which, upon closer examination, prove to be granites and quartz monzonites, intrusive into mica schists, clay slates, altered volcanic tuffs, and amphibolites. The age of the veins is probably early Paleozoic.

Structurally the deposits may be classed as fissure veins and replacement deposits in schists. The veins are in general of the so-called lenticular type illustrated in Fig. 43, in which the quartz

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[1] G. F. Becker, Gold fields of the southern Appalachians, *Sixteenth Ann. Rept.*, U. S. Geol Survey., pt. 3, 1895, pp. 250–331.

H. B. C. Nitze, *Bull.* 10, North Carolina Geol. Survey, 1897.

L. C. Graton and W. Lindgren, Reconnaissance of some gold and tin deposits, etc., *Bull.* 293, U. S. Geol. Survey, 1906.

H. D. McCaskey, Gold, etc., in the Eastern States, *Mineral Resources*, U. S. Geol. Survey, 1908, pp. 645–681 (with literature).

H. D. McCaskey, Notes on some gold deposits of Alabama, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 36–52.

F. B. Laney, The Gold Hill mining district, *Bull.* 21, North Carolina Geol. Survey, 1910.

Stephen Taber, Geology of the gold belt in the James River basin, *Bull.* 7, Virginia Geol. Survey, 1913.

lenses, which collectively constitute the veins, lie parallel to the foliated structure. In detail the lenses often cut across the schistosity and are sometimes of irregular form. Sharply defined veins cutting across the schistosity also occur. The quartz is massive, usually without banded or drusy structure. The replacement deposits form irregular bodies of silicified and pyritic schist; the deposit worked by the Haile gold mine is the most prominent example of this class. The ores form more or less regular shoots, often also pockets, and are in general of low grade; pyritic ore containing $2 per ton has been successfully treated at the Haile mine; many shoots, however, average much higher, sometimes $15 or $20 per ton. Free gold is generally but by no means always present below the zone of oxidation.

Quartz, often glassy and semi-transparent, is the principal gangue mineral and may be accompanied by calcite, dolomite, apatite, chlorite, ilmenite, magnetite, tourmaline, albite and sometimes zinc spinel (gahnite) and garnet. The ore minerals are free gold, pyrite, arsenopyrite, pyrrhotite, molybdenite, more rarely galena, zinc blende, and chalcopyrite. Enargite, tetradymite, altaite, and nagyagite are recorded, but are rare. The pyrite is always the oldest sulphide and the gold fills minute fractures in it, or in the quartz.

The metasomatic alteration of the wall rock shows considerable variations. The most intense action is shown by some quartz-tourmaline veins; the adjoining amphibolite is altered to garnet, tourmaline, and magnetite.[1] In some of the Dahlonega veins the included amphibolite, as well as the adjacent wall rock, is altered to well-developed crystals of pale-red garnet and a dark-green mica. The garnets contain visible gold; the quartz itself contains pyrite, pyrrhotite, galena, and chalcopyrite.[2] This mode of alteration is much like that noted in the rocks adjacent to the pegmatite dikes of the same region.

In other veins a chestnut-brown biotite is the only mineral resulting from metasomatic alteration; in places both muscovite in comparatively large foils and biotite are present, sometimes with calcite or dolomite, besides more or less pyrite or pyrrhotite. Amphibolite is the most easily attacked of the various kinds of country rock. The alteration of granite is usually slight.

The replacement bodies are generally in the acidic schist de-

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[1] L. C. Graton, *op. cit.*, p. 47.

[2] W. Lindgren, *op. cit.*, pp. 126–127.

rived from volcanic fragmental rocks; these are extensively silicified and contain also both sericite and biotite as products of alteration.

Genetically, these gold-bearing veins appear to be connected with granitic intrusions, representing the final product of the most volatile part of the magma. They are considered by Becker, Graton, and Lindgren to be the deepest parts of veins whose upper parts have been carried away by erosion.

The Quartz Veins of Ontario[1]

The gold-bearing quartz veins of Ontario, Canada, are widely distributed. They occur at Lake of the Woods, Rainy Lake, Wahnapitæ, Abitibi, Larder Lake, Kirkland Lake and in the recently discovered Porcupine district. They also extend into Quebec and Manitoba.

Until recently the production of these veins has been disappointing but they have yielded heavily in the last few years owing to the wonderful developments in the Porcupine district. The gold production of Ontario was $10,180,000 in 1916 which exceeds the combined product of British Columbia and Yukon. Almost the whole of this came from the Porcupine.

In some districts, Kirkland Lake for instance, the veins are similar to those of California and contain no distinctive high temperature minerals while at many other places tourmaline and pyrrhotite are found in the ores.

The veins of Ontario are found in the Keewatin greenstone and allied schist, in the later Timiskaming (lower Huronian) conglomerate and greywacke which is infolded with the Keewatin and finally in quartz porphyry, granite and syenite of post-Timiskaming (Algoman) age. The mineralization is caused by these intrusions according to the Ontario geologists. The deposits are

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[1] A. P. Coleman, Reports of Bureau of Mines of Ontario, Nos. 4, 5, 6, and 7, 1894–1896.

W. G. Miller and C. W. Knight, The pre-Cambrian geology of Southeastern Ontario, *Rept.* Ontario Bur. Mines, vol. 22, pt. 2, 1914.

A. G. Burrows, The Porcupine gold area, *idem*, vol. 24, pt. 3, 1915

J. Stansfield, Microscopic examination of Porcupine rocks, etc., Canadian Min. Jour., Feb. 15, 1911.

A. G. Burrows and P. E. Hopkins, The Kirkland lake and Swastika gold areas, *Rept.* Ont. Bur. Mines, vol. 23, pt. 1, 1914.

A. H. Means, Tourmaline bearing gold quartz veins of the Michipicoten district, Ontario, *Econ. Geol.*, vol. 9, 1914, pp. 122–135.

lenticular veins in schist (Fig. 234), fairly regular and branching veins in massive rocks in part also large irregular or dome-shaped masses of quartz. There are scant amounts of simple sulphides such as pyrite, arsenopyrite, pyrrhotite, chalcopyrite, galena, and zinc blende. Molybdenite and tellurides of gold, silver and lead occur in the Kirkland Lake district; scheelite is found at Porcupine. Besides the predominant quartz the gangue minerals include ankerite, dolomite, tourmaline, chlorite, sericite and often also albite. The gold is generally free and often coarse averaging 850 fine. In the Hollinger mine, at Porcupine, the gold occurs in quartz and schist, the latter containing about 5 per cent. pyrite besides sericite, dolomite and chlorite; though probably free, it is so fine that it can not be obtained by amalgamation or panning. Most of the coarse gold appears to be deposited in crushed and fissured quartz, and is somewhat later than the earliest mineralization (Fig. 233).

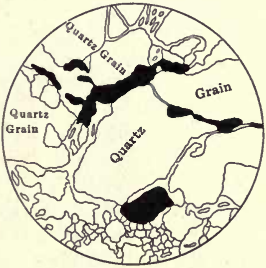

FIG. 233.—Drawing of thin section showing native gold deposited in crushed gold-quartz, Rea vein, Porcupine. Black spots are native gold. Magnified 30 diameters. *After A. G. Burrows.*

At Porcupine the principal mines have attained a depth of 1,250 feet. The Hollinger deposits, which in 1916 yielded about $5,000,000, consist of a series of lenticular veins in schist (Fig. 234). In 1916, 600,000 tons of ore were treated. In 1917 the output decreased somewhat.

Next to the veins the country rock is altered to sericite and carbonates, and carbonatization of basic rocks has often resulted in large masses of coarsely crystalline rock consisting of ankerite and bright-green mariposite (chromium mica). This altered rock, which appears to occupy considerable areas, is often cut by quartz veins and is remarkably similar to certain rocks along the Mother Lode in California, which have resulted from the alteration of serpentine.

In places, for instance, at the Rice Lake veins, Manitoba, biotite is developed in the country rock.

The Pre-Cambrian Gold Veins of the Cordilleran Region

It has already been explained that of the pre-Cambrian veins many, though not all, suggest formation at high temperature. Brief reference suffices to the gold-bearing quartz veins of Hopewell and Bromide,[1] in New Mexico, many of which contain, as metasomatic products, tourmaline, garnet, and other silicates and also the characteristic brown or green mica. Some of the veins occurring in the pre-Cambrian rocks of southern Wyoming,[2] at Atlantic and South Pass, also belong to this type; others give evidence of deposition at lower temperatures.

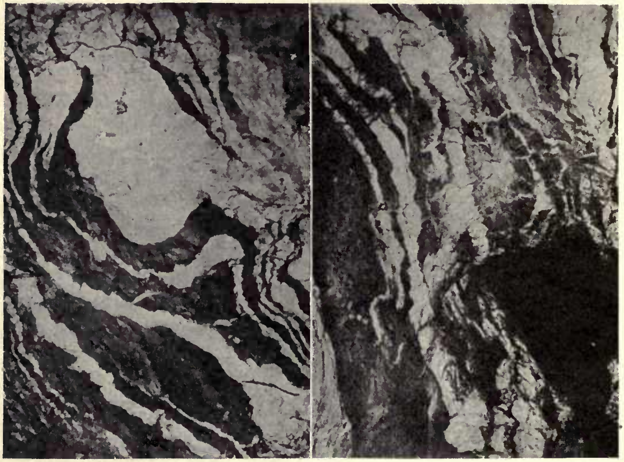
A *B*

FIG. 234, *A* and *B*.—Photographs of veins in Hollinger mine, Ontario. *After A. G. Burrows.*

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[1] L. C. Graton, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 126.

[2] A. C. Spencer, *Bull.* 626, U. S. Geol. Survey, 1916, pp. 9–45.

The Black Hills of South Dakota[1] contain many gold-bearing deposits in the pre-Cambrian rocks. They occur, as a rule, in clay slates of sedimentary origin, and, while some of them are true veins with glassy quartz and free gold, others are lenticular bodies of highly altered rock. The best known among the latter is the Homestake lode at Lead. Embedded in the normal clay slate, which contains not far away great masses of intrusive granite, are huge lenticular bodies of altered rock with quartz, sulphides, and free gold, averaging about $4 per ton. The ratio of silver to gold (by weight) is about 1:5. The numerous ore lenses are in places several hundred feet in width and have been followed for a distance of about 1 mile in the same direction as the strike of the clay slates, which dip steeply to the northeast. The slates and ore-bodies are intruded by rhyolite porphyry, which, however, seems to have caused little if any additional mineralization. At the depth attained, which now is 2,000 feet, the ore-bodies are

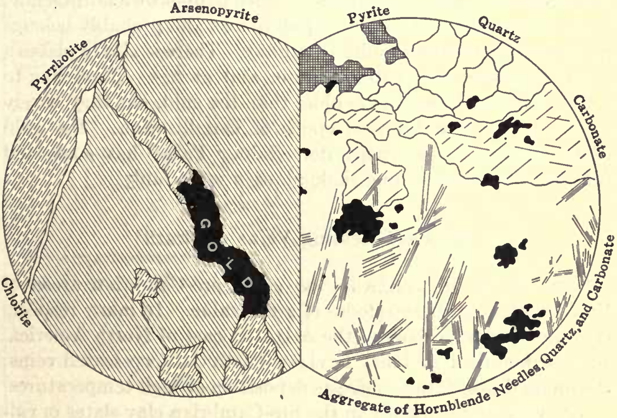

FIG. 235.—Thin sections of Homestake ores. Left: Gold and pyrrhotite (later), in arsenopyrite (earlier). Magnified 48 diameters. Right: Gold with quartz, iron-magnesium carbonate, pyrite, and cummingtonite. Magnified 32 diameters. *After W. J. Sharwood.*

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[1] J. D. Irving and S. F. Emmons, *Prof. Paper* 26, U. S. Geol. Survey, 1904.

W. J. Sharwood, *Econ. Geol.*, vol. 6, 1911, pp. 729–786.

said to maintain their size and value. In a broad way the ore-bodies pitch to the southeast. One thousand stamps crush the ore, and the pulp is amalgamated and afterward leached with cyanide. The production in 1916 had a value of $6,531,000.

The ores differ from the country rock in containing a dissemination of fine-grained free gold, pyrrhotite, pyrite, arsenopyrite, and a little chalcopyrite; the sulphurets are not rich in gold. The ore-bodies also include many small lenticular masses of coarse-grained glassy or milky quartz, which in places contains sulphides but rarely free gold. The larger part of the ore is also distinguished by the appearance of much light-brown hornblende, often with radial structure; it is rich in iron and probably belongs to the species cummingtonite (Fig. 235). There is also in places a little dolomitic carbonate, siderite, and garnet. According to J. D. Irving the silicates are older than the gold, but they surely belong to the same general epoch of metallization. The gold often accompanies arsenopyrite. Sidney Paige[1] has suggested that the ore replaces lime shale along a main fault.

The Gold-Bearing Veins of Brazil[2]

Some of the provinces in the great pre-Cambrian areas of Brazil contain auriferous lodes of great value. In many respects they are similar to those of the Atlantic coast of North America, already described, and may be classed with the deep-seated veins; the mineral association suggests deposition at high temperatures.

The veins occur mainly in the pre-Cambrian clay slates or calcareous slates of Minas Geraes, in southern Brazil, and have been worked successfully to great depths. The mine of St. John del Rey (Fig. 236) has attained 6,300 feet in vertical depth, or 10,000 feet on the incline, and is thus the deepest gold mine in the world. At a depth of 4,900 feet the lenticular ore body is 1,028 feet long

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[1] *Bull.* Geol. Soc. Am., vol. 24, 1913, pp. 293–300.

[2] Orville Derby, Notes on Brazilian gold ores, *Trans.*, Am. Inst. Min. Eng., vol. 33, 1903, pp. 282–287; *Eng. and Min. Journal*, vol. 74, 1902, pp. 142–143.

Georg Berg, Beiträge zur Kenntniss der Goldlagerstätten von Raposos, *Zeitschr. prakt. Geol.*, 1902, pp. 81–84.

E. Hussak, Der Quartzlagergang von Passagem, *Zeitschr. prakt. Geol.*, 1898, pp. 345–357.

Orville A. Derby, Gold-bearing lode of Passagem, *Am. Jour. Sci.*, 4th ser., vol. 32, 1911, pp. 185–190.

and 12½ feet wide. The ore has a value of about $11 per ton. The mineral association consists of native gold, pyrrhotite, pyrite, chalcopyrite, arsenopyrite, quartz, and siderite; albite, apatite, magnetite, and scheelite are also known to occur. The ore contains 28 per cent. pyrrhotite and 25 per cent. quartz, the rest being mainly siderite. Metasomatic replacements often spread laterally into the adjacent schists.

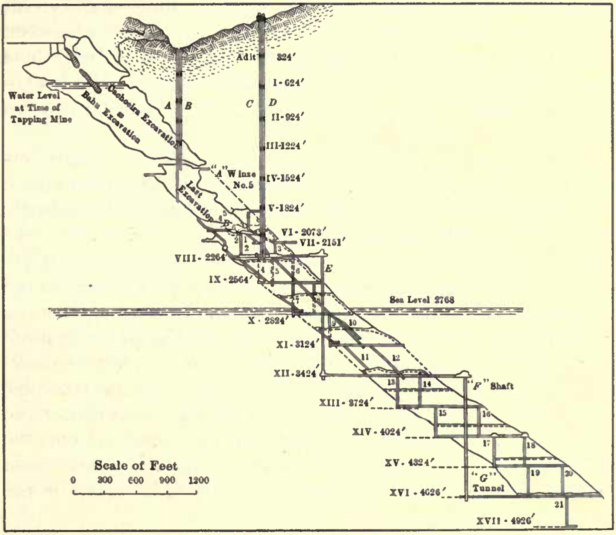

FIG. 236.—Vertical longitudinal section of the St. John del Rey Mining Company's Morro Velho mine, Brazil.

The Passagem lode, described by E. Hussak as a gold-bearing pegmatite dike, appears, according to O. Derby, as a pegmatite dike shattered and impregnated by gold, arsenopyrite, pyrrhotite, and tourmaline, with a little siderite and calcite (p. 776). The gold contains some bismuth. The appearance of cummingtonite allies the Passagem vein in an interesting manner to the Homestake deposit. In other parts of the Brazilian gold-bearing area palladium occurs in alloy with gold.

The Gold-Quartz Deposits of Silver Peak, Nevada

In the Cordilleran region of North America gold-quartz veins of post-Cambrian age belonging to the class of high-temperature veins are not common. Spurr, however, describes such deposits in the Silver Peak district, in western Nevada,[1] which he holds to be closely allied to igneous rocks. Granites with transitions to alaskite and quartz monzonite are here intrusive into Paleozoic limestone. Gold-quartz veins of irregular form and typical alaskite are stated to "form two ends of a rock series between which every gradation is represented." The gold is contained mainly in the pure quartz, which also yields sulphides like arsenopyrite. There is little silver present. The mines closed down in 1915.

The alaskite is a granular aplitic rock containing quartz, orthoclase, microcline, albite, anorthoclase, and oligoclase-albite. The first generation of feldspars was partly altered into muscovite before the second generation of feldspars was deposited. The quartz magma, separated by differentiation, collected in larger masses. Later, repeated mineralization filled the fractures in the magmatic quartz with auriferous pyrite and galena, representing a fresh supply of ascending waters. The hypothesis is advanced that solutions of granitic origin have deposited gold predominantly in the granite or in the rocks silicified by the metamorphic effect of the granite, and that in or near the calcareous rocks more silver and copper were precipitated from the same solutions, the difference being due to the selective influence of wall rock. Garnet and epidote are mentioned as occurring in certain of the veins in calcareous rocks.

The Gold-Quartz Veins of Southeastern Alaska[2]

The gold-bearing veins of southern Alaska are closely allied to those of the Appalachian region and to those of Brazil, although they present some features that would rather connect them with the gold-quartz veins of California, which are believed

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[1] J. E. Spurr, Ore deposits of the Silver Peak quadrangle, *Prof. Paper* 55, U. S. Geol. Survey, 1906.

[2] A. C. Spencer, The Juneau gold belt, Alaska, *Bull.* 278, U. S. Geol. Survey, 1906.

A. Knopf, The Eagle River region, *Bull.* 502, U. S. Geol. Survey, 1912.

to be formed under conditions of lower temperature or more moderate depth.

The veins occur mainly in the narrow strip of sharply folded Paleozoic slates and greenstones which form the western margin of the great batholithic mass of granodiorite of late Mesozoic age, 40 to 80 miles wide and continuous for many hundreds of miles parallel to the coast. The conditions are therefore essentially similar to those of the California gold belt, especially as the Paleozoic sediments farther west on Admiralty Island are adjoined by a belt of slates which are thought to correspond in age to the Mariposa slate of California.

In the long strip of coast country extending 300 miles there are numerous mining districts, among which are Windham Bay, Port Snettisham, Sheep Creek, Gold Creek, Douglas Island, Eagle River, and Berners Bay.

The gold occurs in veins and lodes of various kinds, or more rarely, as on Douglas Island, in altered dikes of dioritic character that contain disseminated free gold and sulphides. The individual veins are rarely continuous for more than a few hundred feet, but often combine to form more extended stringer leads or lode systems. As the veins are later than the schistose structure of the rocks their tendency is to follow foliation planes, and in places they strongly resemble the lenticular veins of the Appalachian region, but continuous and cross-cutting veins also occur.

The gangue minerals are mainly milky quartz with some calcite or dolomite; tourmaline is occasionally reported, also magnetite. The ore consists of free gold containing more or less silver and associated with pyrite, pyrrhotite, zinc blende, chalcopyrite, galena, and arsenopyrite.

Few of the gold-quartz veins have yet been followed to great depth. Their width is from 1 to 8 or 10 feet at most, and the metal content of their ore must necessarily be high, for the amount that can be taken out with profit is small. Their ores would probably range from $5 to $20 or more per ton.

The Treadwell ores, which are mined on a large scale, are of low grade, containing about $3 in gold per ton, of which 60 to 75 per cent. is free-milling, the concentrates yielding $30 to $50 per ton. The Treadwell deposits consist of a series of mineralized dikes (Fig. 237) of albite diorite in slates near the east shore of Douglas Island. The workings extend for 7,000 feet along the shore. The dikes dip about 50° northeast. The dimensions of the dikes are variable, the larger ones having a maximum width of over 200 feet. These ore-bearing dikes have been followed to a depth of 2,400 feet, and there appears to be no diminution of the average tenor of the ore at that depth. The average annual production of the Treadwell mines was, since 1910, $4,000,000. In 1917, an invasion of sea water filled most of the mines.

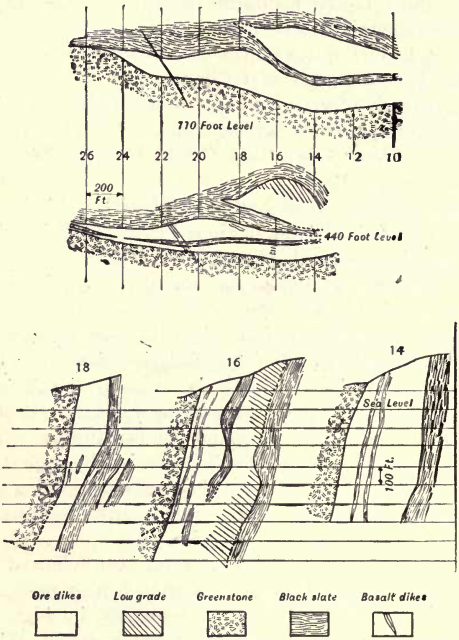

Fig. 237.—Horizontal and vertical sections of ore-bodies in Alaska-Treadwell mine, Douglas Island, Alaska. *After A. C. Spencer.*

The ore-bodies are extensively fractured by a system of conjugated joints, along which there are irregular veinlets of quartz and calcite. The ore minerals are chiefly native gold, pyrite, and pyrrhotite, but chalcopyrite, galena, zinc blende, and molybdenite are also found. The important gangue minerals are albite, calcite, and quartz. The original diorite has been so thoroughly altered that it is difficult to establish its exact character. The metasomatic processes of the Alaska gold-quartz veins, especially the Treadwell dikes, are described below in more detail.

During the last year the great stringer-lodes in slate, like that of the Alaska Gold Mines Company, have been opened on a large scale on the mainland, but with dubious success as the ore scarcely averages $1.25 per ton.

The topographic features of this region permit the generalization that the vertical range of the deposits is over 5,000 feet. They have been followed 2,400 feet below sea level, and typical veins are found in the same regions at elevations of 3,000 feet or more above sea level. They were formed shortly after the great intrusion of granodiorite, and the vertical range now accessible must have been many thousands of feet below the surface of the earth at the time of the ore deposition.

Metasomatic Processes in Veins of Southeastern Alaska

The facts that the Alaska veins contain abundant pyrrhotite and some tourmaline and magnetite and that the altered country rock contains biotite show that in many parts of the region the temperature of deposition was high. Albitization is a common process and appears to be independent of the amount of sodium in the country rock. It takes place not only in albite diorite, where it might be interpreted as a mass reaction, but also in normal diorite, gabbro, and amphibolite.

Adjacent to the crosscutting fissure veins of the Berners Bay district the metasomatic action is, as shown by Knopf, very similar to that in the California gold belt. Dolomite, sericite, albite, and pyrite are the principal new minerals formed in the rock.

The Treadwell mine is working large mineralized dikes of albite diorite in slates and greenstones. According to Spencer[1] the original rock contained albite-oligoclase, microperthite, hornblende, and biotite, the latter two minerals in small amounts. The altered rock contains abundant albite, mostly developed by

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[1] A. C. Spencer, The Juneau gold belt, Alaska, *Bull.* 287, U. S. Geol. Survey, 1906, p. 99.

the replacement of microperthite, also quartz, calcite, muscovite, hornblende, rutile, epidote, magnetite, and pyrite. Albite is also formed as narrow veinlets, although most of the veinlets consist of calcite and quartz. The composition of the altered rock differs considerably from place to place. Spencer holds that sodium has been added to the rock, together with carbon dioxide and sulphur. Calcium in the rock has been fixed by the carbon dioxide and suffered little leaching. The composition of one of the altered rocks is calculated as follows:

Quartz	2.34	Magnesite	0.11
Albite	84.36	Siderite	0.57
Anorthite	1.11	Apatite	0.13
Zoisite	0.91	Rutile	0.13
Muscovite	3.03	Pyrite	2.10
Calcite	3.80		
			98.59

Spencer and Knopf have shown that at several places on the mainland near Juneau dioritic rocks near the veins have been altered to products containing brown mica, probably biotite. Spencer describes the alteration in the Gold Creek district, which results in the development of biotite (I and II in the following table). The mineral composition of the altered rock is calculated as follows: Quartz, 45 per cent.; biotite, 22; carbonates, 20; titaniferous magnetite, 10.5; and sulphides, 2.5.

Knopf describes an altered and fresh amphibolite found near the Mendenhall Glacier (analyses III, IV, V), and calculates the mineral composition approximately as follows:

	Fresh amphibolite	Altered amphibolite
Orthoclase (mol.)	6.7	....
Albite (mol.)	18.3	39.3
Actinolite	43.7	....
Biotite[1]	7.9	43.1
Zoisite	4.5	16.4
Epidote	18.3	....
Apatite	0.6	1.2
	100.0	100.0

[1] By difference.

ANALYSES OF FRESH AND ALTERED ROCKS FROM QUARTZ VEINS OF SOUTHEASTERN ALASKA

(Analysts, George Steiger, I, II, V; J. G. Fairchild, III, IV)

	I	II	III	IV	V
SiO_2	47.76	44.69	48.30	52.92	+ 5.10
TiO_2	1.48	2.25	1.01	0.99	− 0.24
Al_2O_3	13.98	14.97	13.59	20.53	+17.70
Fe_2O_3	1.99	0.60	3.12	Trace	− 9.64
FeO	8.72	7.05	10.44	8.38	− 7.80
MnO	0.14	0.14	0.25	0.28	+ 0.04
MgO	9.07	3.92	6.29	2.43	−12.33
CaO	12.71	10.07	11.09	4.76	−20.00
BaO	Trace	0.14	....	....	....
Na_2O	1.65	2.36	2.16	4.67	+ 6.92
K_2O	0.20	1.76	1.55	2.96	+ 3.83
H_2O-	0.22	0.36	0.00	0.18	+ 0.34
H_2O+	2.06	0.20	2.06	1.58	− 2.25
ZrO_2	....	0.02	....	....	....
CO_2	None	8.47	None	....	....
P_2O_5	0.12	0.26	0.26	0.57	+ 0.86
S	0.04	....	....	....	....
FeS_2	....	0.27	....	....	....
Fe_7S_8	....	2.25	....	....	....
Less O	0.02	....	....	....	....
	100.16	99.78	100.12	100.25	−16.80

I. Green diorite, Gold Creek. Contains about 75 per cent. green hornblende; remainder feldspar with some quartz.

II. "Brown diorite," Ebner mine, Gold Creek.

III. Amphibolite, Mendenhall Glacier. Sp. gr. 3.084. Dark olive-green rock.

IV. Altered amphibolite, Mendenhall Glacier. Sp. gr. 2.905. Dark-brown rock with pyrrhotite.

V. Gains and losses in grams in the alteration of 100 c.c. of amphibolite to same volume of altered product.

These changes differ greatly from those noted along fissure veins of the more ordinary type. In the first place, they include actual dehydration and distinct additions of aluminum, sodium, and potassium, the alkalies having doubled in quantity. In both localities ferric oxide is almost wholly removed, while there is some decrease in the ferrous oxide. Beyond this the two sets of analyses are dissimiliar, for, while one indicates 20 per cent. of carbonates, the other is entirely without carbon dioxide. As a consequence the first has retained much more calcium and magnesium than the second. As to minerals the rock rich in carbonate contains biotite, titaniferous magnetite, and sulphides; the one without carbonates yields albite, biotite, and zoisite, and Knopf considers that apatite has been formed in it.

It is characteristic of the deep-seated veins that actinolite is unstable, while biotite, zoisite, and ilmenite were developed under the influence of the vein-forming solutions. It is believed that the solutions were hot and ascending and that they carried both sodium and potassium, besides phosphorus and sulphur. A similiar development of biotite has been noted in the Kolar gold fields in Mysore, India,[1] which are very productive and are worked to a depth of 4,000 feet. The veins, which are probably of pre-Cambrian age are contained in crystalline schists. The gangue is a glassy quartz with native gold and a small amount of pyrite, pyrrhotite, arsenopyrite, etc. Some tourmaline is present. The veins contain pitching pay shoots in which the ore is 5 feet wide, averaging $20 to $30 per ton in the deepest levels (p. 189).

The Gold-Telluride Veins of Western Australia[2]

Western Australia is an arid tableland of moderate elevation surmounted by short and low ridges (Fig. 238). Crystalline schists and granites are the principal rocks. The schists extend with general north-south strike and vertical or steep dip across the whole central part of the state and consist largely of amphibolites, massive or foliated, which have been derived by metamorphism from a basic rock—diorite, gabbro, or diabase. There

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[1] F. Hatch, The Kolar gold field, *Mem.* Geol. Survey India, vol. 33, pt. 1, 1901.

[2] A. Gibb Maitland, *Bulls.* 4, 15, and 20, Geol. Survey Western Australia.
C. F. V. Jackson, *Bulls.* 13 and 18, *Idem.*
E. S. Simpson, *Bull.* 6, *Idem.*
T. A. Rickard, The telluride ores of Cripple Creek and Kalgoorlie, *Trans.* Am. Inst. Min. Eng., vol. 30, 1901, pp. 708–718.
P. Krusch, *Zeitschr. prakt. Geol.*, 1903, pp. 321–331; 369–378.
C. O. G. Larcombe, The geology of Kalgoorlie, *Proc.* Austral. Inst. Min. Eng., vol. 5, 1910, pp. 1–312.
M. Maclaren and J. A. Thomson, *Min. and Sci. Press*, vol. 107, 1913, pp. 45, 95, 187, 228, 374.

are also highly altered sedimentary rocks such as quartzites and slates; more rarely limestones. At Kalgoorlie, for example, slates are intimately associated with amphibolites.

Granitic rocks, in part gneissoid, also occur extensively in the complex of crystalline schists. Lenticular masses of amphibolite are contained in the granitic rocks and vice versa, so that the sequence of the rocks is not always clearly apparent. Maitland believes that many of the granitic rocks are intrusive into the amphibolites.

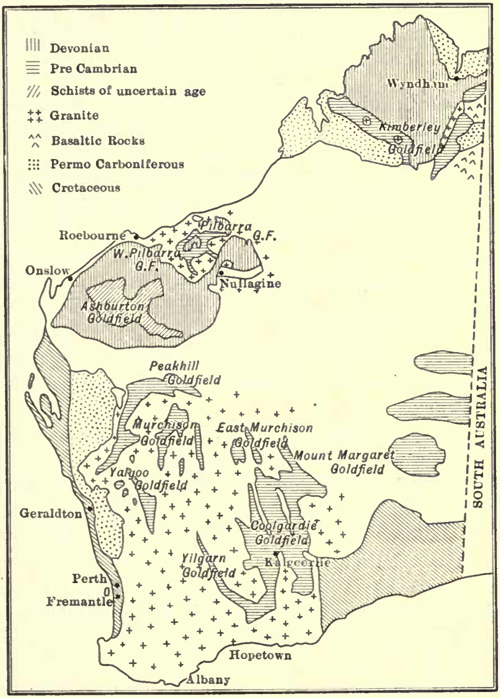

FIG. 238.—Geological map of Western Australia. Scale 1 inch = 330 miles.

The age of the rocks is not definitely known, but is considered pre-Cambrian. Toward the northwest coast the old rocks disappear beneath transgressing horizontal Carboniferous limestones.

The gold deposits are contained chiefly in the amphibolites but also, though less commonly, in the granitic rocks. Maitland says:

> All the important auriferous areas occur within or near the schistose rocks, and they occupy a large area extending from the south coast . . . to the northwest coast, over about 14° of latitude. The auriferous belts exceed 20 miles in width in places.

The center of mining activity is at Kalgoorlie, about 350 miles east of Perth. From the mines within the so-called "Golden Mile" at that place, the larger part of the output of Western Australia has been derived, although other mines north and northwest of Kalgoorlie now contribute a considerable share. The most prominent among these outside mines are the Westralia-Mt. Morgan and Sons of Gwalia, in the Mt. Margaret gold field, and the Great Fingall, in the Murchison gold field, 300 miles northwest of Kalgoorlie. The total gold production of Western Australia from 1886 to 1917, inclusive, is about $660,000,000, the annual yield, which is gradually decreasing, being now about $22,000,000. The yield of the Kalgoorlie mines since discovery is about $330,000,000.

The geologists of Western Australia distinguish two principal modes of occurrence of gold-bearing lodes.

1. The normal quartz veins usually occur in the amphibolite or along the contact of granitic rocks and amphibolite. Most of the veins conform in strike and dip with the steeply dipping schists. The veins are usually short or branched and curved, and the quartz has a tendency to form lenticular ore-bodies. The minerals accompanying the native gold are galena, blende, pyrrhotite, chalcopyrite, arsenopyrite, stibnite, bismuthinite, pyrite, scheelite, chlorite, calcite, and sericite. In addition tourmaline is reported from one mine, the Sons of Gwalia.

Most of the deposits of this class have probably been formed by the filling of open cavities; the veins are often bent, corrugated, and deformed. At Mt. Morgan the quartz bodies form solid pipes of lenticular section, the main axes of which dip 45° to the south. The ore-shoots within these lenses also have a southerly trend.

2. The quartz lenses are at many places surrounded by altered

country rock; where this rock prevails, transitions are formed to the second class of composite replacement deposits or sheared zones, to which the name "lode formations" is given. Simpson describes them as follows:[1]

> A lode formation may be defined as a more or less vertical zone of rock, usually continuous with the surrounding rock and of similar origin, but distinct from it in carrying metallic ores disseminated through it in payable quantities and, as a rule, characterized by strong foliation. The typical lode formations probably owe their origin to a shearing action having crushed and foliated portions of a rock mass in a certain definite direction, producing a more or less well-defined band of rock through which, by virtue of the foliation, mineral-bearing solutions or vapors can have free circulation. In consequence of this, mineral deposits are formed within the rock, usually but not necessarily extending over the whole of the foliated zone, but seldom beyond it, and having no definite boundaries horizontally or vertically other than those determined by the decrease of the assay value of the rock in any one direction.
>
> In the southern portion of the Kalgoorlie belt the rocks consist mainly of amphibolites (altered in various ways, but largely into massive chloritic rock and chlorite schist,) together with some smaller bodies of porphyrite, felsite, graphitic slate, and quartzite. The lode formations consist almost entirely of vertical or steeply inclined zones of chlorite schists or foliated greenstone, often passing insensibly into unaltered greenstone on either side, but sometimes showing an irregular boundary. They vary in width from 2 or 3 feet up to 80 feet.

The minerals of the second type of deposits include native gold and tellurides, such as calaverite, petzite, hessite and coloradoite (telluride of mercury). Pyrite is abundant, but is almost always finely divided, in contrast to the tellurides, which are often massive. Accessory minerals are chalcopyrite, zinc blende, galena, pyrargyrite, enargite, löllingite, fluorite, magnetite, rutile, calcite, dolomite, siderite, ankerite, sericite, chlorite, and roscoelite.[2] To these tourmaline and albite should be added. The ore-bodies, as shown by H. C. Hoover, form lenticular bodies. They have usually a rich core from which the gold content decreases outward, and the lenses are of large dimensions. Mining operations have attained a depth of 3,600 feet, and at this depth some of the mines are still in ore. The chief mines at Kalgoorlie are the Great Boulder, Ivanhoe, Horseshoe, Perseverance, Oroya-Brownhill, Associated, and Lake View Consolidated.

The ore varies from a dark-green, distinctly chloritic foliated

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[1] E. S. Simpson, *Bull.* 6, Geol. Survey Western Australia, 1902, p. 22.

[2] *Bull.* 6, Geol. Survey Western Australia, 1902, p. 21.

schist, as in the Oroya-Brownhill mine, to pale-green sericite schists and to banded or massive dark rocks, flinty in places and ranging from dark green to gray or brown. Small specks of pyrite are distributed through the ores, which contain bright-yellow gold associated with much pale-yellow calaverite and black, lustrous coloradoite with conchoidal fracture. In places the tellurides and gold have developed as seams several centimeters thick, in joint planes crossing the schistosity. The oxidized zone is from a few feet to 200 feet deep. Some geologists believe that secondary tellurides and native gold have enriched the lodes just below the oxidized zone. At greater depths free gold is rarely visible and sulphides tend to take the place of the rich tellurides. The ores have gradually declined in value from $30–$40 per ton near the surface to $7–$10 in the deepest levels. The presence of sodium chloride in the mine waters would suggest much downward transportation of gold.

In Western Australia,[1] as in Alaska, there have been local differences in the processes of alteration. In the Pilbara gold field[2] the granite next to the vein is altered to a greenish-gray rock, the calcium, magnesium, and sodium having been largely removed and the potassium considerably increased. It contains no carbonates. The course of alteration is entirely similar to that of many deposits found at intermediate depths in the Cordilleran region of the United States.

At Kalgoorlie, on the other hand, albite and carbonates are the principal products. The unaltered rock consists of an amphibolite containing amphibole, chlorite, zoisite, and albite. The altered rocks forming the gold-bearing lode contain chlorite, newly formed albite, calcite, dolomite, siderite, tourmaline, sericite, roscoelite, magnetite, specularite, and nests and lenses of fine-grained quartz. The carbonate grains inclose irregular masses of tellurides and coarse gold, but the larger masses of calaverite also contain rhombohedrons of carbonates. Crystals of magnetite embedded in tellurides are reported. E. S. Simpson has shown that the ores are really derived from amphibolites by replacement. The replacement is irregular, albite, quartz, or carbonates alternately predominating.

The character of alteration is shown by the following analyses:

[1] W. Lindgren, Metasomatic processes in the gold deposits of Western Australia, *Econ. Geol.*, vol. 1, 1905, pp. 530–544.

[2] A. Gibb Maitland, *Bull.* 15, Geol. Survey Western Australia, 1904, p. 12.

ANALYSES OF FRESH AND ALTERED AMPHIBOLITES FROM KALGOORLIE

	3231	1936	206	1753	1751
SiO_2	48.86	57.72	51.27	46.94	42.01
Al_2O_3	14.91	9.68	13.85	12.49	8.42
Fe_2O_3	....	6.49	1.54	0.33	2.45
FeO	11.13	9.17	2.63	9.20	15.76
MgO	7.65	1.63	4.18	3.56	1.67
CaO	12.19	5.05	6.40	6.43	7.07
Na_2O	2.58	3.92	1.78	1.84	2.62
K_2O	0.19	0.12	2.37	2.57	1.15
H_2O-	0.04	0.16	0.40	0.09	0.23
H_2O+	1.51	1.51	0.22	0.30	0.67
TiO_2	0.22	1.13	0.23	0.14	0.81
CO_2	None	1.84	8.02	13.41	15.65
P_2O_5	....	....	Trace	....	....
Te	....	....	Trace	....	....
FeS_2	....	....	8.41	2.25	0.30
MnO	0.90	0.09	Trace	0.32	0.41
	100.18	98.51	101.30	99.87	99.22

3231. Star of Colac. *Bull.* 6, Geol. Survey Western Australia, 1902, p. 67. "Rock consists of coarse-grained mixture of feldspar and chlorite. It contains colorless hornblende, saussurite with clear mosaic of albite, also ilmenite surrounded by leucoxene." Analyst, C. C. Williams.

1936. Hannans main shaft, at depth of 600 feet. "Rather coarse grained rock containing hornblende, chlorite, feldspar (albite?), ilmenite, and secondary quartz." Analyst, C. C. Williams.

206. 300-foot level, Lake View Consols mine. *Idem.*, pp. 23 and 67. "Strongly foliated gray lode stuff. Assay, 9 oz. 12 dwt. of gold and 6 oz. 7 dwt. of silver per ton. It shows chlorite and sericite on the cleavage planes." Analyst, E. S. Simpson.

1753. A foliated greenish-gray lode stuff from the 400-foot level, Ivanhoe mine. *Idem.*, pp. 23 and 67. "Contains chlorite, carbonates, a feldspathic material, ilmenite, and some quartz. Trace of gold and no silver." Analyst, C. G. Gibson.

1751. Siderite rock, West crosscut, 400-foot level, Ivanhoe mine. *Idem.*, p. 67. "Gray compact rock, containing carbonates, grains of quartz and of black iron ores, . . . altered feldspar, and some scaly green chlorite." Analyst, C. G. Gibson.

Analysis 3231 is probably fairly representative of many amphibolites of Western Australia. It contains no free quartz and is rich in lime and iron. There is very little potash, and only 2.58 per cent. of soda.

The last three analyses may be roughly calculated as shown in the accompanying table:

PROBABLE MINERAL COMPOSITION OF ORES FROM KALGOORLIE

	206	1753	1751	Remarks
Quartz	29.52	25.20	21.44	
Chlorite	6.86	2.76	2.94	$FeO.4MgO.Al_2O_3.3SiO_2.4H_2O.$
Albite	15.12	15.70	22.12	
Sericite	19.54	21.52	9.58	$K_2O.3Al_2O_3.6SiO_2.2H_2O.$
$CaCO_3$	11.42	11.50	12.61	
$MgCO_3$	4.20	5.38	1.76	
$FeCO_3$	2.32	14.60	23.20	
$MnCO_3$	......	......	0.67	
Pyrite	8.41	2.25	0.30	
Magnetite (?)	......	0.47	3.53	
TiO_2	0.23	0.14	0.81	
Fe_2O_3	1.54	......	......	
Al_2O_3	2.14	0.61	......	
Hygroscopic water	......	......	0.23	
	101.30	100.13	99.19	

At first glance the analyses of the altered rocks do not show very great changes so far as bases and silica are concerned. The principal differences are in the added carbon dioxide of the altered rocks, ranging from 8 to nearly 16 per cent.

Magnesium, calcium and iron have been fixed as carbonates, the latter also as pyrite. The combined water has decreased owing to replacement of chlorite by sericite. The silica set free has been deposited as quartz. Soda has decreased slightly but potash has increased, though not very greatly. The abundant development of albite and carbonates recalls the processes of replacement in the deposits near Angels Camp in California, and at the Alaska-Treadwell mine.

COPPER DEPOSITS

The Copper-Tourmaline Deposits

In several parts of the world the association of chalcopyrite with tourmaline is fairly common. In addition the ores may contain some gold and silver. Other minerals occasionally present are magnetite, specularite, rutile, pyrite, pyrrhotite, molybdenite, bismuthinite, wolframite, scheelite, tetrahedrite, quartz, siderite, fluorite, and biotite. Anhydrite is often present.

The copper-tourmaline veins of Cornwall, which also carry cassiterite, establish the transition from these deposits to the tin-bearing veins. Other occurrences, like those of Meadow Lake, California, and the Passagem, Brazil, emphasize the transition to the gold-bearing quartz veins.

The deposits are in part fissure veins, in part replacements in brecciated or sheared zones. In both modes of occurrence the country rock is subject to intense metasomatic changes, and tourmaline is developed by replacement, often for a considerable distance from the solution ducts. The normal form of alteration in feldspathic rocks is sericitization, sometimes accompanied near the veins by silicification. The feldspars are replaced by sericite, and both feldspar and quartz are penetrated by acicular tourmaline prisms, usually of bluish-gray color. Chalcopyrite, pyrite, and sometimes other sulphides also develop in the altered rock. The final product, unless the tourmalinization is unusually intense, lacks the coarsely crystalline structure of the typical greisen; the mica foils are usually small. In some deposits biotite develops.

Chile.—Many tourmaline-copper deposits are found in Chile. They appear to be connected with basic rocks—gabbro, diabase, porphyrite, diorites, etc.[1]

Von Groddeck has described the formerly important Tamaya mines, where veins containing copper ores cut diabase and porphyrite. The tourmaline is not only present in the filling of the principal vein, which dips 35° and is from 3 to 6 feet thick, but

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[1] A. von Groddeck, *Zeitschr.* Deutsch. geol. Gesell., vol. 39, 1887, pp. 237–266.

A. W. Stelzner, *Zeitschr. prakt. Geol.*, 1897, pp. 41–53.

Moericke, *Tsch. M. u. p. M.*, vol. 12, 1891, pp. 186–198.

L. Darapsky, Das Departement Taltal, Berlin, 1900, pp. 167–172.

is also abundantly developed in the calcitic, chloritic, and micaceous altered country rock. Asbestos and tremolite (cummingtonite) are also mentioned.

Similar veins at Las Condes, 90 miles east of Santiago, have been described by A. W. Stelzner. The rocks are granite and altered andesites. The vein filling consists of pyrite, chalcopyrite, quartz, and a loose mass of tourmaline needles and minute crystals of zircon, octahedrite, and specularite. The country rock is bleached and impregnated with pyrite and tourmaline. At Peralillo, 31 kilometers from Santiago, a similar pyrite-chalcopyrite vein in diorite carries tourmaline, molybdenite, scheelite, and cupro-scheelite.

The most prominent representative of this type in the world is the Teniente, or Braden, deposit, situated in the Western Cordillera, 50 miles S.S.E. of Santiago at an elevation of 8,000 feet. Very large ore-bodies have been developed by tunnels over a vertical interval of 2,500 feet and the production is now about 6,000 tons per day of 2 to 4 per cent. ore.

An intrusive mass of andesite porphyry and monzonite porphyry is injected into Tertiary andesitic lavas and has been extensively mineralized. The greatest ore-bodies surround a volcanic explosion vent about 3,600 feet in diameter. This is filled with rudely stratified tuff and the tuff is again intruded by masses of andesitic breccias of several kinds. The period of greatest mineralization followed the intrusion of the breccias and resulted in the replacement of the rocks surrounding the "crater" by quartz, specularite, tourmaline, chalcopyrite, pyrite and occasionally other sulphides. A still later epoch of mineralization yielded richer ores of chalcopyrite, bornite, tennantite, siderite, rhodochrosite and anhydrite while during the last commercially unimportant epoch were deposited chalcopyrite, bornite, quartz, barite and gypsum. Crystals of gypsum 27 feet long and 2 feet in diameter were found in cavities coated by these products of the last mineralization.

The deposit has been somewhat enriched by chalcocitization effected by descending waters.

The mineralization probably took place at a depth below the surface of about 4,000 feet and was effected by solutions containing much boric acid. They were probably very hot and may have reached the surface in gaseous form like the "soffioni" of Tuscany. The abundance of boron emanations connected with volcanic action in the southern Andes is remarkable. The numerous borax deposits in Chile, Bolivia and the Argentine bear witness that some of these emanations reached the surface.

United States.—In the Cordilleran region of the United States there are a number of similar smaller deposits. They have been described from Meadow Lake, California,[1] from the Blue Mountains of Oregon[2] and from the upper Pecos River in New Mexico.[3]

The most productive deposit of this type occurred at the Cactus mine,[4] in the San Francisco Range, in southern Utah. This deposit is contained in a brecciated zone in post-Paleozoic monzonite; it is in places 200 feet wide and at least 900 feet long. The ores have been followed to a depth of 800 feet. Brown tourmaline, with quartz, pyrite, and chalcopyrite, coats the fragments of brecciated rock, which is sericitized and contains some metasomatically developed tourmaline. Other minerals, formed somewhat later than the tourmaline, are siderite, anhydrite, specularite, and tetrahedrite; these also are associated with chalcopyrite, the development of which continued during the

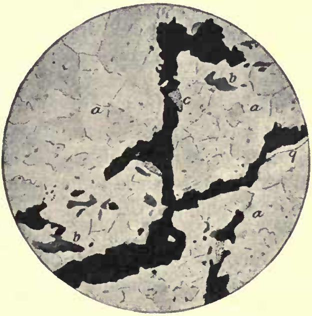

Fig. 239.—Replacement veinlet, War Eagle mine, Rossland, B. C. *a*, Granular aggregate of orthoclase with a little sericite; *b*, biotite; *q*, quartz; *c*, chlorite; black, pyrrhotite. Magnified 40 diameters.

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[1] W. Lindgren, *Am. Jour. Sci.*, 3d ser., vol. 46, 1893, p. 201.

[2] W. Lindgren, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 113.

[3] W. Lindgren, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, pp. 551–776.

[4] W. Lindgren, *Econ. Geol.*, vol. 5, 1910, pp. 522–527.

B. S. Butler, Geology and ore deposits of the San Francisco and adjacent districts, Utah, *Prof. Paper* 80, U. S. Geol. Survey, 1913, pp. 172–178.

whole epoch of mineralization. The ores are of low grade. In 1908, 177,000 tons of ore were mined which yielded copper, 2 per cent.; silver, 0.2 ounce, and gold, 0.01 ounce per ton. The ore was concentrated and smelted. The mine is now closed, poorer ores having been found in depth.

The Gold-Copper Deposits

In a few of these copper deposits tourmaline is absent or rare.

The Rossland district[1] is situated in British Columbia near the boundary line of the State of Washington. It has been producing smelting ores since 1890, and has yielded a total of $62,300,000 in gold, copper and silver. The ores contain about $5 to $10 in gold and 0.3 ounce silver both per ton, as well as 0.5 to 3 per cent. copper. The mines are opened to a greatest depth of 2,200 feet. The deposits are steeply dipping replacement veins (Fig. 239) along shear zones in monzonite and augite porphyrite. Granodiorite appears in some places and is thought to represent upward extensions of the Trail batholith, the emanations from which are believed to have formed the deposits. The ore minerals are chalcopyrite and pyrrhotite with some pyrite, arsenopyrite, molybdenite and bismuthinite. The gangue minerals comprise quartz, magnetite, calcite and biotite with some garnet, arsenopyrite, actinolite and wollastonite. The metasomatic action on the country rock has resulted in much secondary biotite. The veins were formed within the epoch of intrusion for they are intersected by many basic dikes related to camptonite. Apophyllite and other zeolites occur in druses. There is little evidence of secondary enrichment as perhaps is natural in a recently glaciated country.

The great copper mining district of Cobar in western New South Wales[2] contains strong lodes in a deeply eroded desert range of older Paleozoic sediments; they are replacement veins from 10 to 120 feet wide cutting sandstone and slate at acute angles in strike and dip. The lodes show typical "hammock structure" (p. 152) and carry chalcopyrite, magnetite, and pyrrhotite in big lenses. The ores gradually fade into country rock. The

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[1] W. Lindgren, *Trans.* Am. Inst. Min. Eng., vol. 30, 1901, pp. 644–645.

C. W. Drysdale, Geology and ore deposits of Rossland, B. C., *Mem.* 77, Canada Geol. Survey, 1915.

[2] E. C. Andrews, Report on the Cobar copper and gold field, *Mineral Resources* 17, Geol. Survey, N. S. W., 1913.

average contents are 2.5 per cent. copper with $1 to $2 in gold and 2 to 3 ounces of silver per ton. Among the gangue minerals are quartz and an iron silicate, probably ekmannite. There is no arsenic or antimony but some bismuth is present. The greatest depth attained is 1,600 feet. Some lodes with more quartz and less sulphides are worked as gold deposits.

With the water level 200 to 450 feet below the surface and strong, salty mine waters it is not surprising that there was strong enrichment of copper with upper levels as oxidized ores. A short distance below water level secondary sulphides were found, comprising according to Andrews, both chalcocite and chalcopyrite. There was little enrichment of gold and silver.

It is remarkable that no intrusive rocks are found within long distance of these deposits. They were doubtless formed at very great depth and at high temperature.

The Copper-Bearing Veins Allied to Contact-Metamorphic Deposits and Pegmatites

In regions containing contact-metamorphic copper deposits it is not altogether unusual to find pyritic veins which exert an alteration on adjoining limestone similar to contact metamorphism, indicating that the vein-forming solutions possessed a high temperature.

The veins of cupriferous pyrite at Clifton, Arizona,[1] which intersect porphyry and contact-metamorphic limestone, are probably in part of this kind, for it was observed in many places that where they cut across limestone, tremolite and magnetite had developed adjacent to the vein. The primary deposits contain little copper but are enriched by surface waters.

Several occurrences of this kind are reported from New Mexico, particularly from the Sierra Hachita district.

One of the best instances is that of Massa Marittima, in Tuscany, described by B. Lotti[2] and V. Novarese. The great veins carry chalcopyrite, pyrite, galena, and zinc blende and cut across Eocene limestone and clay shales. The limestone, but not the shale, is replaced near the vein by pyroxene, epidote, quartz, and sulphides. Some bismuth and a little tin are present in the

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[1] W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905.

[2] B. Lotti, Descrizione, etc., di Massa Marittima in Toscana, *Mem. descritt. Carta Geol d'Italia*, vol. 8, 1893.

K. Ermisch, *Zeitschr. prakt. Geol.*, 1905, pp. 206–241.

ore. The mineralization is believed to be due to the intrusion of a granite of Tertiary age which on the surface does not come within several miles of the deposit. The whole region, however, gives evidence of strong mineralization.

Copper-Titanium Veins.—The small but interesting group of the chalcopyrite veins associated with titanium minerals is of uncertain affiliations. In some respects they are very closely allied to the pegmatites.

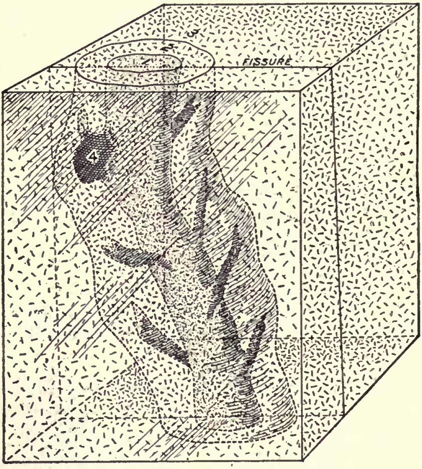

Fig. 240.—Stereogram showing relation of quartz pipe and mineralized quartz monzonite in the O.K. Mine, Beaver Lake district, Utah. 1, quartz; 2, altered monzonite; 3, monzonite; 4, high grade ore. *After B. S. Butler, U. S. Geol. Survey.*

Such deposits have been described from Hereroland in South Africa,[1] at Rehoboth and Otjizongati. They are continuous quartz veins in mica schist and carry pyrite, chalcopyrite, bornite and molybdenite with orthoclase, albite-oligoclase, rutile, ilmenite, apatite and tourmaline. They contain also a little gold.

Copper-Molybdenum Veins.—The association of copper and molybdenum is not uncommon but many other ore minerals are

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[1] Eberhard Rimann, *Zeitschr. prakt. Geol.*, vol. 22, 1914, pp. 223–226.

usually also present. Butler[1] describes the O.K. deposit, Beaver Lake district, Utah, in which the characteristics of pegmatites are curiously mixed with those of high temperature veins. The deposit worked to a depth of 400 feet consists of a cylindrical body of extremely coarse and drusy pegmatitic quartz connected with a steep fissure and surrounded by a zone of sericitized quartz monzonite. The quartz has many offshoots of minor veins, which carry quartz, chalcopyrite and molybdenite (Fig. 240). The ore in the upper levels is oxidized with secondary sulphides.

THE LEAD-SILVER-ZINC DEPOSITS

Veins with Tourmaline.—The combination of galena and tourmaline is rare, galena being generally found in deposits formed at lower temperatures. Recent investigations by A. Knopf,[2] for the U. S. Geological Survey, show that many of the veins in the contact zone and in the igneous rock of the Boulder batholith of quartz monzonite in Montana belong to this unusual group. The Alta vein is the best known and the richest of these deposits; it is supposed to have yielded over $32,000,000 in lead and silver and it was thus one of the greatest lead-silver deposits of the world. The monzonite contains a little tourmaline, its aplite dikes somewhat more, and the quartz veins are rich in this mineral. In the same district H. V. and A. N. Winchell[3] observed a pyrite-tourmaline vein, the ore of which contains mainly silver with some copper and lead minerals. P. Billingsley and J. A. Grimes[4] have also examined these veins and conclude that they have been formed in or near the flat roof of that batholith.

Veins with Garnet.—The great Broken Hill lode[5] in the desert region of western New South Wales is representative of this

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[1] B. S. Butler, *Prof. Paper* 80, U. S. Geol. Survey, 1913, p. 125.

[2] The tourmaline silver-lead type of. ore deposit, *Econ. Geol.*, vol. 8, 1913, pp. 105–118; also *Bull.* 627, U. S. Geol. Survey, 1913.

[3] *Econ. Geol.*, vol. 7, 1912, pp. 287–294.

[4] *Trans.* Am. Inst. Min. Eng., vol. 58, 1918, pp. 284–368.

[5] E. F. Pittman, *Records*, Geol. Survey N. S. W., vol. 3, pt. 2, 1892.

J. B. Jaquet, *Mem.* 5, Geol. Survey N. S. W., 1894.

D. Mawson, *Memoris* Royal Soc. of South Australia, 1912.

R. Beck, *Zeitschr. prakt. Geol.*, 1899, pp. 65–71.

E. S. Moore, *Econ. Geol.*, vol. 11, 1916, pp. 327–348.

See also "Report of Subcommittee," *Trans.* Australas. Inst. Min. Eng., vol. 15, 1911, pp. 160–236.

W. E. Wainright and P. H. Warren, The Broken Hill South mine, *Mining Mag.*, Jan., 1918, pp. 12–19.

rare class. This lode which has yielded lead, silver and zinc to the value of about $450,000,000 since its discovery in 1883, is contained in a folded complex of schists of sedimentary and igneous origin, among which are sillimanite schist, amphibolite, granite gneiss and quartzite, all of probable pre-Cambrian age. The lode has been opened over a length of 3 miles and the deepest shaft has attained 1,800 feet, the ore continuing to that depth.

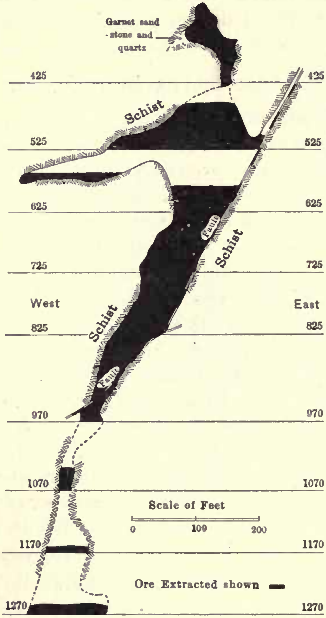

FIG. 241.—Vertical section through the ore-body of Broken Hill South Mine, N. S. W. *After Wainright and Warren.*

A fault zone 6 to 10 feet wide occupies the footwall of the deposit. On the hanging side the ore bulges out in places to great masses, which, as shown in Fig. 241, seem to follow the folded schists and form saddle-like bodies probably formed by replacement of the schist. According to F. E. Wright the quartz in the deposit has been formed at temperatures below 575° C. The ore contains galena and zinc blende, with subordinate pyrite; the gangue includes manganese, garnet, rhodonite, quartz and calcite. The ores contain from 3 to 14 ounces of silver per ton, 14 to 16 per cent. lead and 8 to 18 per cent. zinc.

The surface gave little indication of the character of the deposit. Down to a depth of 300 feet there was a gossan, 20 to 100 feet wide of quartz, limonite, manganese dioxide, hematite and kaolin. Below this were found great masses of cerussite, anglesite, cuprite and malachite, with abundant cerargyrite, embolite and iodyrite. There was but little smithsonite, the zinc having been removed by leaching.

Where the oxidized ores changed to primary sulphides there was a thin deposit of sooty chalcocite, rich in silver and copper; the slight depth of these secondary sulphides is remarkable.

The important deposits in the Kootenay district of British Columbia must have been formed under similar conditions. According to S. J. Schofield[1] they may be considered as high temperature equivalents of the Coeur d'Alene lead deposits.

The ores form replacement deposits of argillaceous quartzite of pre-Cambrian (Algonkian) age. These rocks probably rest on intrusive granite.

The ore-bodies conform roughly to strike and dip of the quartzites; the greatest dimensions are 825 by 120 feet. The ore is an intimate mixture of galena and zinc blende with minor amount of pyrite, pyrrhotite, magnetite and jamesonite; the scant gangue contains red garnet, diopside, actinolite and biotite with subordinate calcite. The gangue minerals are earlier than the sulphides.

THE COBALT-TOURMALINE VEINS

The association of tourmaline with nickel and cobalt minerals in San Juan, Department Freirina, Chile, has been described by O. Stutzer.[2] In the same paper he gives a general review of the tourmaline veins.

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[1] *Econ. Geol.*, vol. 7, 1912, pp. 351–363.

Kootenay District, B. C., *Summ. Rept.*, Canada Geol. Survey, 1915, pp. 93–94.

[2] *Zeitschr. prakt. Geol.*, 1906, pp. 294–298.

CHAPTER XXVII

DEPOSITS FORMED BY PROCESSES OF IGNEOUS METAMORPHISM

INTRODUCTION

General Features.—In many geological provinces and during all ages molten magmas have invaded older rocks without reaching the surface. The intrusive magma cooled slowly and crystallized either as rocks with coarsely granular texture, such as granite, diorite, syenite, monzonite, gabbro, or diabase, or as the corresponding porphyries with holocrystalline groundmass. By means of uplift and subsequent erosion, these igneous rocks become exposed at the surface. If the rocks bordering the intrusives are crystalline schists or older igneous rocks, they seldom show much alteration along the contacts, but where they are of sedimentary origin, like sandstone, shale, and limestone, considerable metamorphism is effected in them for a varying distance from the contact. In many places deposits of metallic ores or other useful minerals occur at these contacts, particularly where the older rock consists of limestone.

The form of such deposits is irregular and bunchy, but many of them are tabular by reason of following the contact (Fig. 242) or certain strata in the intruded rocks favorable for deposition (Fig. 243). Their mode of occurrence is almost wholly metasomatic—that is, they are formed by replacement of the enclosing rock.

The mineral association is characteristic: Chalcopyrite, pyrite, pyrrhotite, zinc blende, and molybdenite are the most common sulphides; magnetite and specularite the most common oxides. The most prominent gangue minerals are various silicates of calcium, magnesium, iron, and aluminum, which have been partially furnished by the carbonate rocks and shales. Among these so-called contact-metamorphic silicates are garnet, epidote, vesuvianite, diopside, tremolite, and wollastonite. Recrystallized, sometimes exceedingly coarse calcite is abundant; quartz is rarely present in large amounts. The ore minerals are usually later than the silicates.

Some of these deposits contain valuable non-metallic minerals, like graphite or corundum, but ordinarily they are mined for the base metals. In the main the ore and minerals are of simple composition and formulas. The association indicates an origin at high temperature, perhaps from 300° to 600° C. Close to the igneous rock the temperature may have been materially higher.

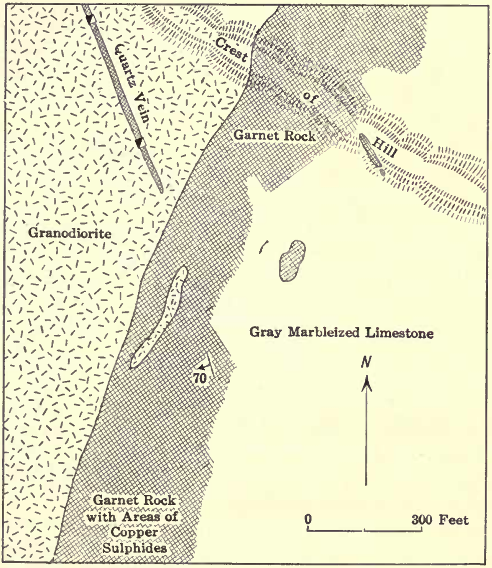

Fig. 242.—Sketch showing relation of ore zone to granodiorite and limestone, Bullion district, Nevada. *After W. H. Emmons.*

The igneous rock itself may be wholly fresh or it may contain minerals closely allied to those in the deposit itself, such as garnet and epidote, or it may show sericitization with veins somewhat later than the alteration at the contact.

History.—In 1865 Bernard von Cotta described the iron deposits of the Banat province of Hungary and expressed the opinion that they were due to the action of intrusive rocks on the adjoining Mesozoic limestone. He also correlated these ores with those of Bogoslowsk, in the Ural Mountains, of Kristiania, in Norway, and of other districts. Von Groddeck,[1] however, first recognized them as a definite group to which he gave the name "Kristiania type." He stated that they were produced by contact metamorphism and called them briefly "contact deposits." Some of the examples mentioned by von Groddeck and others are doubtful, and in later text-books the type was rather neglected.

In the last years of the nineteenth century Vogt[2] revived the interest in this class by describing the contact-metamorphic deposits of Kristiania. A little later the deposits at Seven Devils, Idaho[3]—the first of this type to be noted in the United States—were described, and in a paper on the character and genesis of certain contact deposits[4] the type was redefined and a number of examples from the United States were cited. W. P. Blake[5] mentioned the frequent occurrence of this type in Arizona. W. H. Weed[6] described a number of additional sub-types; a

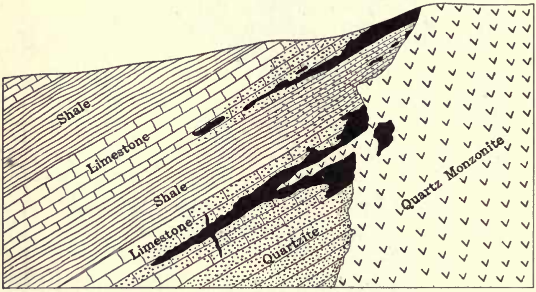

FIG. 243.—Diagram of contact-metamorphic deposit in vertical section. Ore shown in black. Contact-metamorphic rocks beyond ore stippled.

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[1] A. von Groddeck, Die Lehre von den Lagerstätten der Erze, Leipzig, 1879, p. 260.

[2] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1894, pp. 177, 464; 1895, p. 154.

[3] W. Lindgren, *Min. and Sci. Press*, vol. 78, 1899, p. 125.

W. Lindgren, *Twentieth Ann. Rept.*, U. S. Geol. Survey, 1900, pt. 3, pp. 249, 253.

[4] W. Lindgren, *Trans.*, Am. Inst. Min. Eng., vol. 31, 1901, p. 23.

[5] W. P. Blake, *Trans.*, Am. Inst. Min. Eng., vol. 34, 1904, pp. 886–890.

[6] W. H. Weed, Ore deposits near igneous contacts, *Trans.*, Am. Inst. Min. Eng., vol. 33, 1903, pp. 719 *et seq.*

little later J. F. Kemp,[1] O. Stutzer[2] and J. E. Spurr[3] discussed the subject again from a general standpoint; W Lindgren, J. F. Kemp, S. F. Emmons, J. E. Spurr, and others described in detail similar deposits in Arizona, New Mexico, and Mexico, and it became evident that this type was far more common than had been suspected It was found that in many regions intrusive masses were normally accompanied by contact-metamorphic deposits which in some cases were connected by transitions with the swarm of veins that usually surround these igneous bodies as an aureole of metallic treasure. The great importance of this type for the solution of problems related to the genesis of ore deposits became clear to the minds of many investigators. In Europe many geologists have of late made detailed studies of contact-metamorphic deposits—among them B. Lotti, R. Beck, Loewinson-Lessing, E. Weinschenk, A. Bergeat, and V. M. Goldschmidt.

The views expressed in the above-mentioned papers, involving accession of material from the magma have not been allowed to pass unchallenged. F. Klockmann[4] expressed the opinion that these deposits were older accumulations of iron ore, altered at the intrusive contact; W. O. Crosby[5] and A. C. Lawson[6] held that such bodies were simply the result of the ordinary circulation of meteoric character. As a general explanation none of these views appear to be tenable.

In 1914 the discussion[7] regarding the origin of the garnet zones flared up again and was participated in by W. L. Uglow, W. Lindgren, J. F. Kemp, C. K. Leith, A. C. Lawson, and C. A. Stewart.

CONTACT METAMORPHISM

General Features.—It will first be necessary to enter a little more deeply into the problem of contact metamorphism. This peculiar action of intrusive igneous bodies upon adjacent sedi-

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[1] J. F. Kemp, Ore deposits at the contact of intrusive rocks and limestone, *Econ. Geol.*, vol. 2, 1907, pp. 1–13.

[2] O. Stutzer, Kontaktmetamorphe Erzlagerstätten, *Zeitschr. prakt. Geol.*, vol. 17, 1909, pp. 145–155.

[3] A theory of ore deposition, *Econ. Geol.*, vol. 7, 1912, pp. 485–492.

[4] F. Klockmann, *Zeitschr. prakt. Geol.*, 1904, pp. 73–85.

[5] W. O. Crosby, *Trans.*, Am. Inst. Min. Eng., vol. 36, 1906, pp. 626–646.

[6] *Min. and Sci. Press*, Feb. 3, 1912.

[7] See *Econ. Geol.*, vols. 8 and 9; *Trans.*, Am. Inst. Min. Eng., vol. 48, 1915, and *Min. and Sci. Press*, Oct. 17, 1914.

mentary rocks has been a well-known fact in geology since the days of Durocher (1846), and the processes have been described in much detail. Effusive rocks—that is, lava flows—rarely exert intense metamorphism beyond a baking or hardening of the sediments at the contact or an alteration of included rock fragments. The magmas intruded in sedimentary rocks, on the other hand, are in most cases surrounded by a halo of gradually fading metamorphism which may extend over a width of 1 or 2 miles, although usually much narrower. The immediate contact is ordinarily sharp, with no evidence of melting; only at contacts which were deeply submerged is there evidence of assimilation and extensive injection or feldspathization of the sediments. Slates and shales are ordinarily converted to hard, compact "hornfels"—that is, fine-grained holocrystalline rocks containing biotite, andalusite, staurolite, scapolite, garnet, and feldspar; in extreme cases gneissoid rocks result. This metamorphism gradually diminishes and at some distance the only evidence of change is a knotty texture of the rocks. Sandstones change to quartzite at the contact. Calcareous rocks become highly crystalline marbles and usually develop the contact-metamorphic minerals garnet, epidote, diopside, tremolite, vesuvianite, etc. According to the older view expressed by Rosenbusch, Zirkel, Brögger, and others, there is little change in composition aside from the expulsion of carbon dioxide from the limestones. The silicates are held to be formed under the influence of the heat of the magma from the impurities contained in the limestone.

It is well known that Rosenbusch[1] proposed a way of calculating the original character of a metamorphic rock from its present composition, and in regional metamorphism this is undoubtedly often justified. More recently this thought has been followed by J. Barrell[2] in more direct application to contact metamorphism.

G. W. Hawes,[3] however, many years ago pointed out that emanations of boron and silica entered the sediments from the magma, and the introduction of tourmaline, for instance, has often been proved since then; it is in fact admitted even in the older text-books of petrography.[4]

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[1] Elemente der Gesteinslehre, 2d ed., 1901, p. 484.

[2] J. Barrell, The physical effects of contact metamorphism, *Am. Jour. Sci.*, 4th ser., vol. 13, 1902, pp. 279–296.

[3] G. W. Hawes, *Am. Jour. Sci.*, 3d ser., vol. 21, 1881, p. 21.

[4] F. Zirkel, Lehrbuch der Petrographie, Leipzig, vol. 2, 1894, p. 97.

The development of greisen by the action of fluorine vapors on granite is in some of these books regarded as a contact-metamorphic process, but in almost all cases it is really a distinctly later stage. J. S. Flett,[1] for instance, has shown that in Cornwall both tourmaline and topaz, where present on a large scale, have been introduced into rocks already contact-metamorphosed.

Petrographers of the present day, represented, for instance, by Alfred Harker and A. Lacroix, lean strongly toward the view that "it is safe to assume that the water and other volatile constituents actually found in igneous rocks, whether chemically combined or mechanically enclosed, represent only a fraction of what was originally contained in the parent rock magmas. The rest has been lost and in the case of intrusive rocks must have passed into or through the surrounding country rocks. Probably some leakage goes on throughout the process of consolidation, but not with equal freedom at different stages . . . and it is likely that a large part of the volatile constituents is in general retained down to a late stage. Nevertheless, more or less of the water and other gases must pass into the neighboring rocks while these are still heated by the intrusion."[2]

The surrounding rocks of some intrusive bodies were, for instance, permeated for long distances by chlorine solutions, this action resulting in the development of scapolite ($Na_4Al_3Si_9O_{24}Cl$). Just how far the substances liberated from the magma are in the state of perfect gases—that is, above the critical temperature—or in liquid form, is not easy to decide. At the contact the carbonate rocks appear to have been permeable to gases to a remarkable degree.

If we admit that the original magma contained in solution various volatile substances, such as water, carbon dioxide, sulphur, boron, chlorine, and fluorine, it follows that the decrease of pressure caused by its ascent will result in the escape of a large part of these volatile compounds, which will entrain with them various metals also held in solution. The higher the rise of the magma the more complete will be the liberation of these substances. In what form, time, and quantity they will escape depends upon pressure, chemical affinities, and miscibility.

The escape of these substances, as pointed out by Harker, may

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[1] The geology of Bodmin and St. Austell, *Mem.*, Geol. Survey England, Explan. Sheet 347, 1909, p. 58.

[2] Alfred Harker, The natural history of igneous rocks, 1909, pp. 302–303.

not have been uniform. A large part was doubtless given off while the magma was still fluid. Another part may have been liberated at the time of consolidation; still another part may have been retained until cooling had advanced considerably. Finally fissures and shattering of the partly consolidated or congealed mass may have permitted gases from the still fluid interior or basal part to reach the outside of the intrusions.

It does not seem possible that atmospheric waters could have gained access to the contacts during intrusion. Both the heat of the magma and the pressure of the volatile compounds striving to free themselves would prevent such access.

The magmatic solutions enter into the rocks adjacent to the magma and produce a series of metasomatic changes the character of which will depend upon the composition of the solutions and their temperature, as well as upon the kind of country rock exposed to these hot extracts.

Ore deposits of this kind are rarely formed in argillaceous shales, sandstones and quartzite. Limestone, dolomite and calcareous shale on the other hand are easily permeable and appear to soak up the solutions like a sponge. It will be recalled that even at ordinary temperature limestone will absorb oil and similar substances. This permeation is facilitated by the fissuring and crushing along the contact. If the solutions are dilute little replacement occurs but the limestone usually becomes coarsely crystalline or scattered silicates will be produced by reaction between the calcite and the silica usually contained in most limestones. When the solutions were more concentrated and contained much silica, sulphur, iron and other metallic constituents the replacement will proceed with energy and the calcareous rock may be converted to a mass of ore and gangue minerals. In places whole beds of pure limestone or dolomite may be converted to garnet, diopside and other silicates associated with recrystallized calcite and with magnetite, specularite and simple metallic sulphides.

Form and Texture.—The frequency of rudely tabular bodies dependent upon parallelism with contacts or beds has already been emphasized. Sometimes the outlines are entirely irregular, but a selective action is very common by which the development of the best ore takes place along certain horizons in which the limestone is especially susceptible to metamorphism. This is doubtless caused by the physical condition of a particular bed which may be a pure or an impure limestone, but the reason for this selective action is not always clearly apparent. While the deposits rarely extend far from the contacts cases are known where they reach out along certain beds for a distance of 2,000 feet or more. The contact-metamorphic minerals in limestone often cease suddenly and form sharp contacts with the unaltered rock. Fissures may guide the solutions and transitions to veins may be formed by deposition along them.

Favorite points of maximum ore development are limestone fragments included in the igneous rock or points where the calcareous rock projects into the intrusive. In most places the ore-bodies are of comparatively small size, but occasionally we find ore-bodies containing millions of tons. In case of copper deposits, subsequent oxidation and enrichment has often, as at Bisbee, Arizona, greatly enlarged the available ore-bodies.

The texture of the ore is generally coarse, characteristic of replacement at high temperature. Fine-grained ores like those of Bisbee are not common. The absence of banding by crustification or deposition in open space is notable. But replacement processes sometimes result in a rude banding by a rhythmic metasomatic action or in orbicular or pipe-like arrangement of minerals. Such orbicular structures have been described by Trüstedt[1] in tin deposits at Pitkäranta, Finland and by Knopf[2] in similar deposits in Alaska.

The calcite is often recrystallized to extremely coarse masses. Garnet and other silicates frequently show crystal outlines but the sulphides, excepting pyrite, are rarely crystallized in distinctive form. Drusy cavities into which quartz crystals, garnets, etc., project are found in some deposits.

Mineralogy.—Among the ore minerals those of simple composition prevail. In order of abundance we have pyrite, chalcopyrite, bornite, pyrrhotite, zinc blende, molybdenite, arsenopyrite and galena. Tetrahedrite, jamesonite and other complex sulphosalts are rare; most of those common in deposits formed at lower temperature are absent. Tellurides are rare though tetradymite is reported from a few localities. Native gold is seldom present though most of the ores contain a trace of that metal. Graphite is not uncommon and platinum has been reported from one or

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[1] O. Trüstedt, Die Erzlagerstätten von Pitkäranta, *Bull.* Com. Geol. Finlande, vol. 19, 1907.

[2] Adolph Knopf, *Bull.* 358, U. S. Geol. Survey, 1908.

two localities. The oxides are represented by magnetite, ilmenite, hematite, quartz, corundum, cassiterite and spinels of various kinds.

The silicates, especially those containing lime, magnesia and iron, are abundant. We enumerate garnet (grossularite and andradite) epidote, zoisite, diopside, hedenbergite, tremolite, vesuvianite, forsterite, anorthite, ilvaite and wollastonite. Some of these contain the hydroxyl molecule.

Among the minerals containing chlorine, fluorine or boron we mention scapolite, chondrodite, axinite, ludwigite, topaz and tourmaline. The latter two are not abundant.

Other minerals occasionally present are orthoclase, albite, muscovite, biotite, wolframite, scheelite and fluorite while barite and anhydrite are absent. Andalusite, cyanite and staurolite, though common in contact-metamorphosed shales, are not usually connected with ore deposits. Chlorite is sometimes found in large crystals; serpentine is a product of decomposition of other silicates. Calcite, ankerite and dolomite are usually present. Many zeolites are also found but always crystallize during the last phase of mineralization.

On the whole, iron and copper deposits are the most common. Zinc, lead, gold, silver and tin are much less abundantly represented.

Intensity of Metamorphism.—There are all degrees of metamorphism at the contacts of intrusive masses. Some magmas are evidently poor in volatile constituents and may exert only a slight recrystallization on adjacent limestone, while along the contact of others large bodies of ore may form. Along the same contact there are often great irregularities in mineralization. The degree of alteration of non-calcareous shales is a good indication of the intensity of the metamorphism, though these rocks rarely contain ore deposits.

Influence of Composition of Igneous Rock.—Highly acidic rocks, such as normal granites, are not usually accompanied by ore deposits of the contact-metamorphic type, although they may produce widespread effects of metamorphism and a later mineralization of quartz veins. The rocks accompanying the contact-metamorphic deposits of the Cordilleran type are generally monzonites or quartz monzonites or granodiorites.

Many examples show, however, that more basic rocks also may produce metallization of adjoining limestones—diabases, for instance, at Cornwall, Pennsylvania, and gabbro at the Nickel Plate mine, British Columbia.

Alteration of the Intrusive Rock.—Fresh granitic rock often adjoins the contact and is then separated sharply from the altered limestone. In many cases the intrusive rock is more or less altered; the alteration may have taken place simultaneously with the metamorphism of the sedimentary rock, or afterward.

The hot intrusive, whether molten or just consolidated, and the adjacent cooler limestone, form a system in which by means of gases, there will usually take place a vigorous exchange of material. Most of the changes will take place at the cooler side but the intrusive will also receive material perhaps mainly carbon dioxide, and oxides of calcium and magnesium from the sediments. Thus we find frequently epidote in large masses, more rarely garnet, or diopside developing in the igneous rock by replacement through the mass or in replacement veinlets. Such phenomena have been described frequently by Spurr and Garrey from Dolores[1] and Velardena,[2] by Bergeat[3] from Concepcion del Oro, all in Mexico, and by Umpleby[4] from White Knob, Idaho. At the latter place it is often impossible to define the contact between garnetized limestone and granite-porphyry (Fig. 244).

In other cases the alteration observed consists in sericitization and impregnation with pyrite. It is believed that this is largely caused by hot waters similar to those which form ordinary fissure veins; these waters may ascend from deeper portions of the intrusive from which emanations continue to be given off for considerable time after the irruption.

In the ordinary course of events an intrusive suffers various changes as to composition and texture at a contact. Basic facies may appear, or fine-grained texture, or a laminated or schistose structure.

The history of any igneous rock is marked by a successive development of minerals. Flett[5] states that in the Cornwall

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[1] J. E. Spurr and G. H. Garrey, *Econ. Geol.*, vol. 3, 1908, pp. 688–725.

[2] J. E. Spurr, G. H. Garrey and C. N. Fenner, *Econ. Geol.*, vol. 7, 1912, pp. 444–492.

[3] A. Bergeat, *Boletin* 27, Inst. Geologico de Mexico, 1910.

[4] J. B. Umpleby, *Prof. Paper* 97, U. S. Geol. Survey, 1917.

[5] J. S. Flett, *Mem.*, Geol. Survey England, Explan. Sheet 347, Bodwin and St. Austell, 1909, p. 58.

granites tourmaline, white mica and topaz may be primary minerals and have a long period of formation continuing to develop until the quartz separates out, and even long afterward.

Spurr and Garrey state that at Dolores and Velardeña many changes took place after consolidation in the small stocks of diorite, alaskite and monzonite breaking through Cretaceous limestone, and that a transition between contact-metamorphic deposits and normal veins are found at the former place. Pegmatitic vuggy segregations are observed and the older minerals have been altered to an early phase characterized by orthoclase, apatite, titanite, chlorite, quartz, diopside and pyrite. At Velardena grossularite, garnet and vesuvianite also appear.

This is followed at Dolores by iron silicates, such as hedenbergite (iron pyroxene) and andradite, both replacing earlier minerals. After this deposition of garnet, but overlapping it, were formed pyrite, chalcopyrite, actinolite, fluorite and quartz, and this association also occurs in the fissure veins. Zoisite, prehnite and apophyllite occur in the later phases, which like the former were formed by solutions rising through the already congealed monzonite.

In the surrounding limestone the succession of minerals is garnet (oldest), magnetite, hematite, pyrite, chalcopyrite, zinc blende and galena.

Succession of Events.—The idea of successive epochs during the intrusion is frequently advanced. The general metamorphism by heat and dilute vapors is generally considered to have taken place first while foreign substances were added later. However, if we conceive magma rising to regions of lessening pressure it is difficult to see why the escape of gases should not have begun at once when the melt was brought into contact with the cooler, surrounding sediments. The frequency of contact-metamorphic deposits in roof segments of intrusive masses and along dikes in the roof shows that the volatile substances were concentrated in the upper parts or cupolas of the intrusives and in dikes radiating from these "gassed" parts of the magma. We observe also the scant metamorphism along flat intrusions, like laccoliths and its relative strength in crosscutting bodies where gases from below could more easily gain access.

The description of a few typical localities will show that the contact phenomena may vary greatly in different places. At times the general metamorphism is very weak while the metasomatic phases are strong and vice versa. Undoubtedly emanations along fissures continued after the general consolidation in many places and if they had a high temperature the later alteration would be of the same type as that produced during the intrusive act.

In an admirable study of contact metamorphism at Marysville, Montana, Barrell[1] describes the metamorphic zone, ½ to 1 mile wide surrounding a stock of quartz monzonite which on the surface occupies an area of only 2 square miles, but which widens below. The intensity of the metamorphism is thus increased.

Barrell distinguishes between (1) contact metamorphism, which results in recrystallization of the sediments to hornfels, marble and lime silicate rock, and this was produced by the first wave of metamorphism; (2) contact metasomatism in which magmatic emanations added some constituents to the altered rocks. This latter zone is at most 1,000 feet wide. Silica, iron and sulphur were added. Diopside and hornblende, with a little apatite, tourmaline, garnet and pyrite were the minerals formed. There are no contact-metamorphic deposits of economic importance.

A very different state of affairs is described by Calkins[2] from Philipsburg, Montana, where a larger batholith invades Algonkian sediments of all kinds. The contact zones are half a mile or more in width. The author draws no sharp line between metamorphism and metasomatism, and evidently considers both to progress at the same time. The metamorphism is strong; close to the contacts are masses of magnetite, in part carrying gold, with garnet, vesuvianite, humite and forsterite. On the other hand scapolite and tourmaline are distributed in the sediments for a distance of a mile or more from the contacts showing a widespread diffusion of magmatic chlorine and boron. Fluorine has also been introduced, further sodium, silica and iron, the latter only close to the contact.

A monograph by V. M. Goldschmidt[3] describes in great detail the type locality of the Norwegian contact-metamorphic deposits near Kristiania. At present they are of little economic impor-

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[1] Joseph Barrell, *Prof. Paper* 57, U. S. Geol. Survey, 1907.

[2] W. H. Emmons and F. C. Calkins, *Prof. Paper* 78, U. S. Geol. Survey 1913.

[3] Die Contactmetamorphose im Kristianiagebiet, Kristiania, 1911.

tance. On a basement of Archean rocks rest Paleozoic sediments; these are broken by laccoliths of gradually more acidic composition, beginning with essexite, which is followed by syenitic rocks. Near the contacts of the essexite the metamorphism is exceedingly strong but takes place without addition of substance.

Along the syenite contacts Goldschmidt observed both an older normal contact metamorphism by recrystallization without addition and a younger "pneumatolytic" metamorphism by recrystallization under the addition of magmatic gases, resulting in "skarn rocks,"[1] in which the copper deposits are contained. The characteristic minerals of the hornfels or altered slates were formed before the consolidation of the magma and probably without the aid of other water than that normally contained in the rock. Though the "skarn rocks" and their metallic sulphides are later than the general metamorphism they were formed shortly before the crystallization of the magma, though the immediate contact may have been congealed. Pyroxene occurs in the inner and amphibole in the outer contact zone; according to Becke the transition point between the stability fields of the two minerals is about 550° C. at 200 atmospheres.

The "skarn rocks" are coarsely crystalline and consist of andradite, hedenbergite, wollastonite, scapolite, axinite, adularia, albite, calcite, fluorite, zeolites, specularite, magnetite, bismuthinite, galena, chalcopyrite, primary chalcocite, primary willemite, zinc blende, pyrrhotite, molybdenite, and bornite. Many of these also occur as primary minerals in miarolitic cavities in the syenite. Magnetite forms nearer to the igneous rock than specularite. The scapolite becomes unstable at lower temperatures and is transformed to albite, epidote, microperthite and zeolites. The metallic ores are somewhat later than the skarn minerals.

From Ely, Nevada, Spencer[2] describes contact zones along small stocks of quartz monzonite. These zones extend a few hundred feet to half a mile from the contact. The monzonite porphyry is greatly altered with development of sericite, biotite and pyrite but no garnet or diopside. The process involved loss of sodium, calcium and some alumina; gain of potassium and sulphur.

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[1] An old Swedish mining term signifying the garnet-pyroxene-epidote rocks accompanying many Scandinavian magnetite deposits.

[2] A. C. Spencer, *Prof. Paper* 96, U. S. Geol. Survey, 1917.

The shales are converted to strongly pyritic hornfels involving addition of sulphur and iron. The limestones have developed white and brown mica, tremolite, pyroxene, garnet, epidote, scapolite, pyrite, pyrrhotite, chalcopyrite, galena, zinc blende, molybdenite, magnetite and hematite. In places very heavy masses of jasperoid have been formed from limestone. In general, silicon, sulphur, potassium iron, copper and other metals have been added.

Spencer again draws no sharp line between metamorphism and metasomatism, conceiving that the whole process took place by emanations from deeper parts of the magma after the parts now at the surface had consolidated. The areas exposed he considers too small to have effected the changes shown. It seems probable that in this case the alteration of the porphyry and the silicification of the limestone resulted from a somewhat later mineralization though practically continuous with the earlier phase.

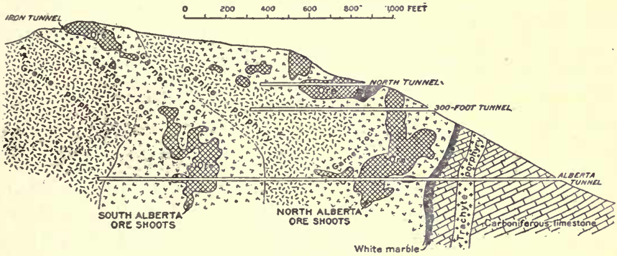

FIG. 244.—Longitudinal section, Empire mine, White Knob, Idaho. *After J. B. Umpleby, U. S. Geol. Survey.*

Very different conditions exist at White Knob, Idaho, described by Umpleby.[1] Here the Carboniferous limestone is very little altered in contact with fresh granite porphyry. Marble has developed close to the contact and more extensively in a roof segment of the batholith with scattered crystals of wollastonite, tremolite and diopside. Engulfed blocks of limestone are, however, more extensively garnetized with development of considerable ore shoots (Fig. 244). The process shows great additions of iron, alumina and silica. To some extent the granite

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[1] J. B. Umpleby, *Prof. Paper* 97, U. S. Geol. Survey, 1917.

porphyry is also garnetized so that in places the contacts are indistinct, and it contains also diopside which replaces biotite and hornblende. The feldspars are the last minerals of the intrusive to be affected. Two stages of metamorphism are recognized: (1) Contact metamorphism at time of intrusion; (2) contact metasomatism after the consolidation of the magma. The garnet and the ore was developed during the last stage. That the last process took place after the intrusive had solidified is proved by garnetization following joint planes in the rock.

In the San Francisco district, Utah,[1] Butler finds silicate zones one-fourth mile wide and recrystallization of limestone over much wider areas. The contacts of garnetized limestone and fresh monzonite are often exceedingly sharp, the transition zone being less than one inch in width.

All this shows that the process varies considerably in different magmas, and that there are considerable differences of opinion as to the various stages of the process.

Succession of Minerals.—The introduction of sulphides and other metallic ores into the limestone is too obvious to be disputed. Many observers have noted a certain succession of developments. Generally, the silicates and the magnetite are earlier than the sulphides, but the periods of deposition overlap.

In the deposits at White Horse, Northwest Territory, Stutzer[2] reports the succession: wollastonite, pyroxene, magnetite, garnet, calcite and sulphides; at Berggieshübel, Saxony, pyroxene, garnet, pyritic ores, zinc blende, arsenopyrite. At the Holgol Mine, Korea, Koto[3] found the succession to be ilvaite,[4] diopside, garnet, while the sulphides fill the interstices between the diopsides (Fig. 251). Frequently chalcopyrite replacing calcite cements the garnet and other silicates (Fig. 245).

In the Boundary district,[5] British Columbia, magnetite, garnet and epidote were formed together while pyrite followed by chalcopyrite came later though partly overlapping in sequence.

Other authors admit that the sulphides are in part later but point out that to a considerable extent they have crystallized

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[1] B. S. Butler, *Prof. Paper* 78, U. S. Geol. Survey, 1913.

[2] *Zeitschr. prakt. Geol.*, vol. 17, 1909, pp. 116–120.

[3] *Jour. Coll. Sci.*, Imp. Univ., Tokyo, May 28, 1910.

[4] Hedenbergite according to D. F. Higgins, *Econ. Geol.*, vol. 13, 1918, p. 19.

[5] O. E. LeRoy, *Mem.* 21, Canada Geol. Survey, 1912.

together with the silicates. This seems to be the condition in the White Knob, San Francisco, Ely, Clifton[1] and Camp Hedley[2] districts. At the Imperial Mine, Utah, Butler finds magnetite in part later than chalcopyrite.

Takeo Kato[3] found that at the Okufo Mine, Japan, the succession began by wollastonite, which is replaced by andradite. The sulphides are deposited contemporaneously with the andradite or at the very close of its deposition.

C. H. Clapp[4] describes contact-metamorphic deposits on Vancouver Island, in which the order is diopside, epidote, garnet, magnetite, pyrrhotite, pyrite and chalcopyrite but the silicates continued to form after some metallization had taken place.

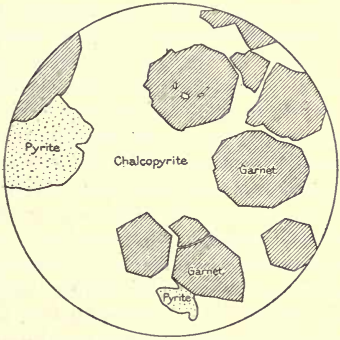

FIG. 245.—Garnet crystals in matrix of chalcopyrite replacing residual calcite. *After J. B. Umpleby, U. S. Geol. Survey.*

The succession of the sulphides generally begins with arsenopyrite and pyrite; then follows pyrrhotite, chalcopyrite, galena, zinc blende; the rarer sulphosalts were formed last.

The early appearance of pyrite is explained by Spencer[5] by the accession of additional iron in the solutions from the sediments and igneous rock which would bring about precipitation of iron compounds before those of copper. He regards the

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[1] W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905.

[2] C. Camsell, *Mem.* 2, Canada Geol. Survey, 1910.

[3] *Jour. Geol. Soc.*, Tokyo, vol. 20, 1913, pp. 13–32.

[4] *Mem.* 13, Canada Geol. Survey, 1912, p. 158.

[5] A. C. Spencer, *Prof. Paper* 96, U. S. Geol. Survey, 1917, pp. 64–72.

magmatic solutions to contain much SiO_2, H_2S, KSH, CO_2, HCO_3, F, Fe, and Cu. The relative abundance of metals in the deposits at Ely is Fe, Cu, Pb, Zn, Mo and in the main that is also the sequence of deposition. This leads to the suggestion that the principal control in determining the order of deposition of the sulphides was the relative concentration of metal radicles in the mineralizing solutions.

Volume Relations.—The replacement of limestone by sulphides, oxides, and silicates liberated a large volume of carbon dioxide, and this at first probably was above the critical temperature; possibly a portion may have been resorbed in the magma, but a larger part was doubtless dissipated in the fractures surrounding the intrusive mass and gradually escaped or mingled with the escaping magmatic water and some distance away with the ground waters, thus adding to the load of ascending thermal springs.

In the Morenci district, Arizona, the clearest evidence is given by the transformation effected along a dike of unaltered quartz monzonite porphyry, 20 to 50 feet wide, which crosses the successive Paleozoic formations with no evidence of fractures that could have admitted solutions from the depths after the consolidation. In the lower limestones the contact zones are only a few feet wide, consisting of epidote next to the intrusive rock, followed by garnet, which adjoins the unaltered limestone. The addition of iron and silica to this narrow zone, which shows no evidence whatever of contraction of volume, is so clear that it hardly admits of discussion. Farther up the same dike cuts across a pure limestone about 80 feet in thickness. This has been changed to massive andradite garnet, with some epidote, for a distance of about 100 feet from the dike. Stains of malachite are present, but this particular rock is poor in copper.

If all of the lime has been used in the garnetization and only CO_2 has escaped, the volume of the rock would have increased about one-half. If, on the other hand, as seems probable, the volume has remained approximately constant, then 460 kilograms CaO and 1,190 kilograms CO_2 per cubic meter have been carried away, while 1,330 kilograms SiO_2 and 1,180 kilograms Fe_2O_3 have been added. These are astonishing figures and give an idea of the vigorous transfer of material which took place during metamorphism. The Modoc limestone contains 94 to 96 per cent. $CaCO_3$, less than 1 per cent. $MgCO_3$, 1 per cent.

SiO_2, and 1 to 3 per cent. Al_2O_3 and Fe_2O_3. The andradite garnet contains 1.53 per cent. alumina, 31.41 Fe_2O_3, 42.63 SiO_2, and 23.37 CaO. The transfer has been mutual, for at some places the intrusive rock next to the contact has been strongly epidotized by lime derived from the calcareous rock.

That the volume has remained practically constant even in the most intense metasomatism may be considered proved and confirmed by the observations in the Ely, San Francisco and White Knob district referred to above. Preservation of structure like stratification planes, joints and fossils has repeatedly been observed in the silicate rock and in the sulphides.[1]

C. K. Leith and E. C. Harder[2] have attempted to account for the silicate rock at the contact of the Iron Springs intrusive in southern Utah by assuming that the silica and alumina have remained constant while the volume of the rock has been decreased as much as 60 or 80 per cent. by reason of abstraction of lime. No definite field evidence, however, was found in favor of this view.

Calkins[3] thinks that at Philipsburg a part but not the whole of the shrinkage was offset by accessions chiefly of silica and alkalies, and that the shrinkage may have been obliterated by pressure.

Mode of Transfer.—During the alteration of the carbonate rock much of the carbon dioxide, lime and magnesia[4] was carried away. In various proportions this has been compensated by additions from magmatic emanations of silica, iron, alumina, sodium and perhaps potassium, and a number of other useful metals. Mineralizers like sulphur, chlorine, boron, fluorine and arsenic have also been introduced. Many writers, like Leith and Harder, Calkins and Goldschmidt, believe that the magmatic gases consisted largely of chlorides, fluorides, etc., and this view is very likely correct.

The equations roughly representing these transformations in case of iron would be:

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[1] Adolph Knopf, *Bull.* 580, U. S. Geol. Survey, 1915, pp. 1–18.
W. Lindgren, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 294.

[2] *Bull.* 338, U. S. Geol. Survey, 1908. See also review by Kemp, *Econ. Geol.*, vol. 4, 1909, p. 782, and answer by Leith, *Econ. Geol.*, vol. 5, 1910, p. 188.

[3] *Prof. Paper* 78, U. S. Geol. Survey, 1913, p. 132.

[4] J. M. Boutwell observed a concentration of magnesia in the altered limestone at Bingham, Utah, *Prof. Paper* 48, U. S. Geol. Survey, 1905.

$$2FeF_3+3CaCO_3=Fe_2O_3+3CaF_2+3CO_2$$
$$2FeCl_3+3CaCO_3=Fe_2O_3+3CaCl_2+3CO_2$$

Physical Conditions at the Contact.—The temperature at the contact, according to the composition of the magma, may have been as much as 1,500° C., the siliceous rocks consolidating at a temperature of 500° to 1,100° C. When there is no chemical action involved calcium carbonate begins to lose carbon dioxide at 550° C., but the reaction would begin at a much lower temperature under the influence of magmatic gases acting chemically on the calcite. Under abyssal conditions no opportunity would be afforded for the liberated carbon dioxide to escape through the fissured rocks.

Even where the carbon dioxide can not escape there may be intense action between the igneous rock and the limestone. The two rocks will form a chemical system with great difference of temperature and it may be assumed that there will be intense transfer of material between the two. Possibly lime and carbon dioxide will be absorbed by the magma in exchange for metallic constituents exhaled from the igneous rock. At great depths the action will be sustained over a longer period and the results may be somewhat different from those obtained within the cooler and brittle upper zone.

A gradual lowering of the temperature from the maximum obtaining at the contact will take place 1. during the diffusion of the gases outward into cooler sediments; 2. during the gradual cooling of the intrusive itself. The presence of wollastonite indicates that the temperature at this particular place could not have exceeded 1,300° C. above which point this mineral is unstable.

Lindgren and Whitehead[1] attempted to determine the temperature by the solubility curve of sodium chloride, which salt is present in sharp cubes in fluid inclusions in quartz in a contact-metamorphic deposit at Zimapan, Mexico. They concluded that the temperature of formation was about 400° to 500° C.

Wright and Larsen[2] have shown that the quartz in contact-metamorphic deposits was formed below 575° C. Spencer[3] examined fluid inclusions in jasperoid at Ely, Nevada, and by

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[1] *Econ. Geol.*, vol. 9, 1914, pp. 435–462.

[2] *Am. Jour. Sci.*, 4th ser., vol. 27, 1909, pp. 421–447.

[3] *Prof. Paper* 96, U. S. Geol. Survey, 1917, p. 63.

calculations based on relative volume of gas and liquid found probable temperatures of 200 ° to 350°C. The jasperoids were, however, probably formed at lower temperatures than the contact-metamorphic silicates.

Depth of Formation.—In many cases it is, of course, difficult to ascertain the depth below the surface at which contact-metamorphic deposits were formed. In the province which contains the most abundant and charcateristic examples of this type, however, namely, the Cordilleran region of America, the conditions of sedimentation and intrusion were such that approximately correct measurements are feasible. Brögger, many years ago, pointed out that granular intrusive rocks by no means always crystallized at abyssal depths and that some intrusions in the Kristiania region had congealed much less than 1,000 feet below the surface.

In the central Cordilleran region sedimentation was almost continuous from the Cambrian to the late Cretaceous; the intrusive rocks now exposed were injected into these sediments during Cretaceous or earlier Tertiary time, well up in the zone of fracture and far above normal "anamorphic" conditions. In 1907 Barrell[1] showed that at Marysville, Montana, the batholith reached within 4,000 feet of the surface, and Leith and Harder gave the same figure for the iron deposit at Iron Springs, Utah. In New Mexico[2] similar conditions existed. At the close of the Cretaceous period practically the whole State was covered by a mantle of sedimentary rocks from 6,000 to 9,000 feet thick. The Cretaceous part of this section, into which most of the numerous intrusive masses reached, was between 3,000 and 5,000 feet thick; much of it consisted of tough but pliable shales not easily broken through by the intrusions. At many places contact-metamorphic deposits were formed less than 3,000 feet below the surface. C. R. Keyes[3] arrived at similar conclusions.

The intrusive "trap" sheets of Triassic age in Connecticut, New Jersey, and Pennsylvania have exerted some contact-metamorphic action and produced small copper deposits; at Cornwall, Pennsylvania,[4] important magnetite deposits were formed in

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[1] *Prof. Paper* 57, U. S. Geol Survey, 1907.

[2] W. Lindgren, L. C. Graton, and C. H. Gordon, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 41.

[3] *Econ. Geol.*, vol. 4, 1909, pp. 365–372.

[4] A. C. Spencer, *Bull.* 359, U. S. Geol. Survey, 1908.

calcareous Carboniferous rocks. The depth below the surface was probably less than 1,000 feet.

For contact-metamorphic deposits in pre-Cambrian areas and in general where periods of dynamic metamorphism have intervened exact data of this kind can rarely be obtained. Some deposits of this class were formed at great depth and under distinctly abyssal conditions.

Piezo-Metamorphism.—Where the intrusion and metamorphism took place under dynamic conditions—that is, under strong pressure from one direction—the results may be expected to differ from those already described. Such dynamic conditions did not exist in the Cordilleran region during the Cretaceous and Tertiary intrusions, but would be more likely to occur in the abyssal zone. Probably many enigmatic deposits of the pre-Cambrian have been formed in this manner.

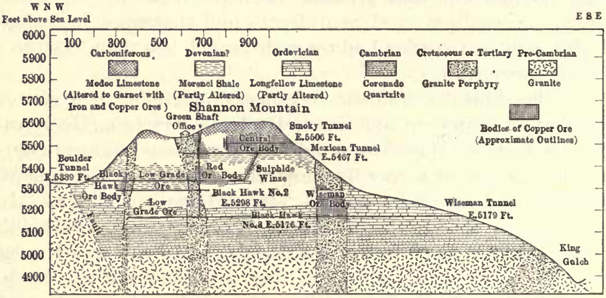

FIG. 246.—Vertical section showing flat ore-bodies at Shannon mine, Clifton, Arizona. Ore-body in porphyry dike consists of secondary chalcocite.

E. Weinschenk[1] has studied this kind of alteration in the Alpine region and names it piezo-metamorphism. According to him many of the gneisses of the central Alps are post-Carboniferous intrusives, pressed during metamorphism. Adjoining limestones have been made crystalline and contain characteristically rounded crystals of quartz, corundum, and micaceous and chloritic minerals.

In the northern part of the Cordilleran region the deposits seem to be less abundant, though several representatives may be

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[1] E Weinschenk Allgemeine Gesteinskunde, 1902

found in Canada, Alaska, Montana, Idaho, Nevada, and Utah. There are relatively few of them in Colorado, but they occur in much greater abundance in New Mexico, Arizona, and Mexico.

A recent reconnaissance of the metal deposits in New Mexico permits a good review of the frequency and relationships of these ores. Along a belt extending from the northern boundary down to the southwestern part of the State the Paleozoic and Mesozoic strata are intruded by at least 20 stocks of early Tertiary quartz monzonite or monzonite, usually of moderate size. The major part of the commercial mineral deposits cluster around these intrusions. Contact-metamorphic deposits were found in 14 districts and at least 25 mines have been worked on a commercial scale. At San Pedro and Jarilla primary chalcopyrite ores are smelted; at Magdalena the deposits yield zinc, copper, and lead; at Hanover, magnetite and chalcopyrite. In the minor deposits the ores may simply form irregular masses at the contact, rarely extending more than 200 feet away from it.

In southern Arizona the deposits are equally common. Among them are the copper deposits of Clifton, Bisbee, Saddle Mountain, Twin Buttes, Washington, Silver Bell, Ajo, Casa Grande, and Vekol. At Clifton and Bisbee the ores have been greatly enriched by oxidation; at Saddle Mountain, Twin Buttes, Washington, and Silver Bell primary chalcopyrite ores are worked.

In eastern Mexico cupriferous contact-metamorphic deposits are common where monzonites break through Mesozoic limestones, as first mentioned by Ordoñez and Aguilera. Since then detailed descriptions have been given of the districts of Santa Fe, in Chiapas; Velardeña, in Durango; San Jose, in Tamaulipas; Concepcion del Oro, in Zacatecas; Dolores, in San Luis Potosi; and Cananea, in Sonora.

PRINCIPAL TYPES OF CONTACT-METAMORPHIC DEPOSITS

The contact-metamorphic deposits may be classified as follows:

1. Magnetite deposits.
2. Chalcopyrite deposits. Principal ore minerals are chalcopyrite, pyrite, pyrrhotite, zinc blende, molybdenite, magnetite, and specularite.
3. Galena and zinc blende deposits.
4. Arsenopyrite-gold deposits. Principal minerals are arsenopyrite and pyrrhotite.

5. Gold deposits.
6. Cassiterite deposits.
7. Titanium deposits.
8. Scheelite deposits.
9. Graphite deposits.

The chalcopyrite deposits present the most common type; the magnetite deposits are fairly abundant, while the ores containing galena, arsenopyrite, gold, or cassiterite are distinctly rare.

Magnetite Deposits

General Character.—The magnetite deposits of this class are of common occurrence, though rarely very large. Associated with the magnetite is more or less specularite, almost always a little pyrite and chalcopyrite, and the contact silicates andradite, ilvaite, olivine, and hedenbergite—all four rich in iron (Fig. 247). The magnetite is sometimes crystallized and often developed in coarsely granular masses (Fig. 248). The ore-bodies are of irregular form, unless, as often happens, they follow the stratification for some distance.

Foreign Occurrences.—Among the European deposits those of Berggiesshübel, in Saxony; Schmiedeberg, in Silesia; and Gora Magnitnaja and Wyssokaja Gora, in the Ural Mountains, are usually described in the text-books. Regarding the latter two occurrences, the opinions seem to be somewhat divided.

The classical deposits of the Banat province, in southeastern Hungary, first described by von Cotta, deserve special mention. In this region early Tertiary intrusive rocks, designated as diorite, syenite, and their porphyries, break through Mesozoic limestones. Along the contacts the limestones become coarsely crystalline, and the usual metamorphic silicates,[1] together with irregular masses of magnetite and some sulphides, develop in them. A banded structure, sometimes apparent, is caused by alternating layers of garnet and magnetite of contemporaneous origin. Masses of garnet from 70 to several hundred feet thick occur. The best-known mines of this region are at Moravicza, Dognacska, and Oravicza. The present annual production is only about 150,000 short tons. According to Bergeat, there can be no doubt that the ores are of contact-metamorphic origin.

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[1] A ferromagnesian borate, ludwigite, is recorded from Moravicza

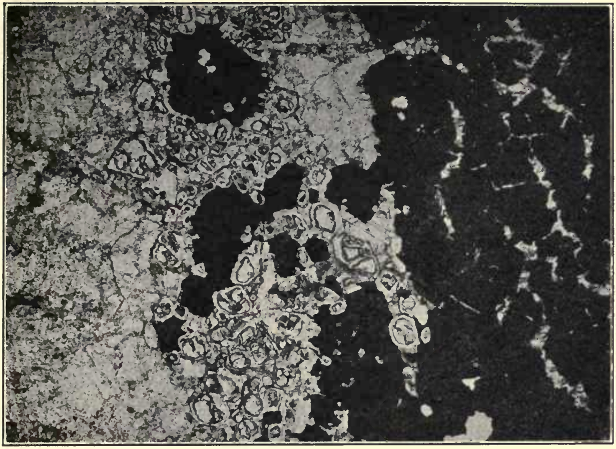

FIG. 247.—Thin section showing magnetite replacing limestone in contact-metamorphic zone, Philipsburg, Montana. Intermediate zone rich in olivine. *After F. C. Calkins. U. S. Geol. Survey.*

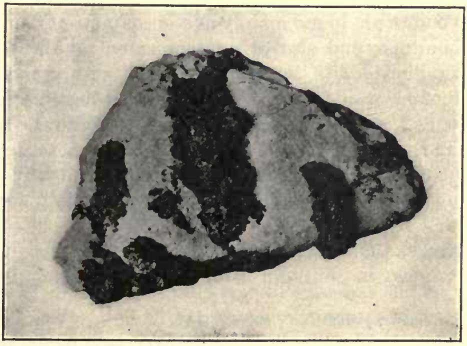

FIG. 248.—Magnetite replacing limestone in contact-metamorphic zone, Cable mine, Philipsburg, Montana. Natural size. *After F. C. Calkins.*

The celebrated mineral deposits of Elba,[1] Italy, with their beautifully crystallized hematite, are likewise of contact-metamorphic origin and were formed under the influence of post-Eocene granite.

Fierro, New Mexico.—Many magnetite deposits of this kind are known in the United States, particularly in the Western States, but most of them are comparatively small. A deposit at Fierro,[2] in southwestern New Mexico, is actively worked at present, the ore being shipped to Pueblo, Colorado. The ore occurs at the contact of quartz monzonite porphyry, probably of early Tertiary age, with Paleozoic limestone; it outcrops in bold masses and is mined in open cuts. The ore-bodies are mainly irregular, lenticular masses of magnetite with a little chalcopyrite; in part they are pure magnetite containing from 60 to 70 per cent. of iron. Those parts which contain a notable quantity of chalcopyrite are left as pillars. Small bunches of garnet and epidote are present in the ore, and in places there are horses of more or less metamorphosed limestones; the phosphorus is rarely above 0.07 per cent.; the sulphur averages 0.02 per cent. Similar deposits, richer in copper, have been mined for flux, the ores being used in the copper furnaces at Douglas, Arizona.

Heroult, California.—Another deposit is situated in Shasta County, California.[3] The ore is smelted locally at Heroult, in an electric furnace.

The ore-bodies are found mainly at the contacts of diorite and Triassic limestone, and also to a minor extent at the contacts of the same diorite with Permian shale and with granodiorite. The order of crystallization appears to have been as follows: 1. Magnetite; 2. garnet and hedenbergite; 3. ilvaite and quartz; 4. pyrite and chalcopyrite. The limestone is practically pure, and that material has been transferred from the intrusive seems to be the unavoidable conclusion. The ore is low in phosphorus and sulphur.

Iron Springs, Utah.—The important, yet unworked deposits of Iron Springs, in southern Utah, have been described by

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[1] B. Lotti, *Zeitschr. prakt. Geol.*, 1905, p. 141.

[2] L. C. Graton, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 313.
Sidney Paige, *Bull.* 380, U. S. Geol. Survey, 1909, pp. 199–214.
Sidney Paige, *Folio* 199, U. S. Geol. Survey, 1916.

[3] Basil Prescott, *Econ. Geol.*, vol. 3, 1908, pp. 465–480.

C. K. Leith and E. C. Harder.[1] A laccolith of quartz syenite porphyry (andesite according to the nomenclature of the authors) breaks through sediments of Carboniferous, Cretaceous, and Tertiary age (Fig. 249). The magnetite appears in fissure deposits and replacements along the contact with the Carboniferous limestone. Quartz, garnet, diopside, apatite, and hornblende are minor constituents of the ore. According to the authors only a part of the ore is associated with contact metamorphism, for the probable gaseous emanation of iron compounds continued after the consolidation, and the resulting magnetite, sometimes associated with apatite, garnet, etc., filled contraction fissures in the intrusive and replaced the limestone near the contact.

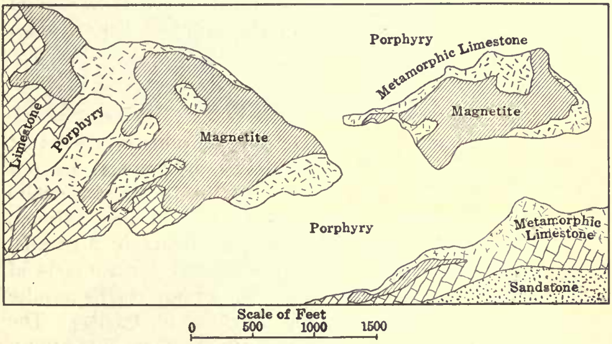

FIG. 249.—Plan showing magnetite and limestone in projecting point of limestone, Iron Mountain, Iron Springs, Utah. *After Leith and Harder.*

But this by no means proves that magnetite was not also generally introduced on a large scale during the early metamorphic action. In fact, most observers of contact metamorphism agree that magnetite is introduced at an early stage, generally before the sulphides.

Harder[2] describes certain vein-like masses of magnetite and hematite, associated with garnet and epidote, in granite of San Bernardino County, California.

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[1] *Bull.* 338, U. S. Geol. Survey, 1908.

[2] *Bull.* 430, U. S. Geol. Survey, 1910, pp. 228–239.

Cornwall, Pennsylvania.[1]—The Carboniferous consists at Cornwall of shales and sandstone, with beds of impure limestone, and is intruded by sheets of diabase probably consolidated at slight depth below the surface. Near the contact the shales are somewhat baked, and in the calcareous sandstone some garnet, pyroxene, quartz, and pyrite have developed. The limestone, near the contact, is replaced by irregular masses of magnetite, accompanied by garnet, pyroxene, epidote, albite, pyrite, and chalcopyrite. In some places the ores, which are mined on a fairly extensive scale, continue along the slight dip of the beds for several hundred feet from the contact. The minerals were formed in the following order: Garnet (not abundant), pyrite, pyroxene, magnetite, feldspar, epidote. Some chalcopyrite is present and is later than the magnetite. Both Spencer and Harder believe that the iron in the magnetite was derived from the molten diabase magma.

Chalcopyrite Deposits

General Character.—The contact-metamorphic deposits that carry chalcopyrite as the predominating ore mineral are not abundantly represented in Europe, but are the most common type in North America, particularly in New Mexico, Arizona, and Mexico. Similar deposits have been found in Australia, Japan and Korea. They occur as a rule at the contacts of smaller intrusive masses of monzonite or quartz monzonite against limestone. Their form is irregular or tabular. The tabular deposits follow certain beds in the limestone formations, and their hanging and foot walls may consist of little-altered or unaltered limestone. The structure of the ore is massive and coarse granular (Figs. 250 and 251). The ore minerals consist of chalcopyrite, bornite, pyrite, more rarely pyrrhotite, and zinc blende, often also molybdenite and other sulphides; galena is on the whole rare. The ore contains also more or less magnetite and specularite. The gangue minerals are andradite, grossularite, epidote, diopside, tremolite, ilvaite, and calcite. The deposits are poor in gold and silver and are frequently enriched

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[1] A. C. Spencer, Magnetite deposits of the Cornwall type in Pennsylvania. *Bull.* 359, U. S. Geol. Survey, 1908, pp. 74–76. *Idem, Bull.* 430, 1910, pp. 247–249.

E. C. Harder, Structure and origin of the magnetite deposits near Dillsburg, Pennsylvania, *Econ. Geol.*, vol. 5, 1910, pp. 599–622.

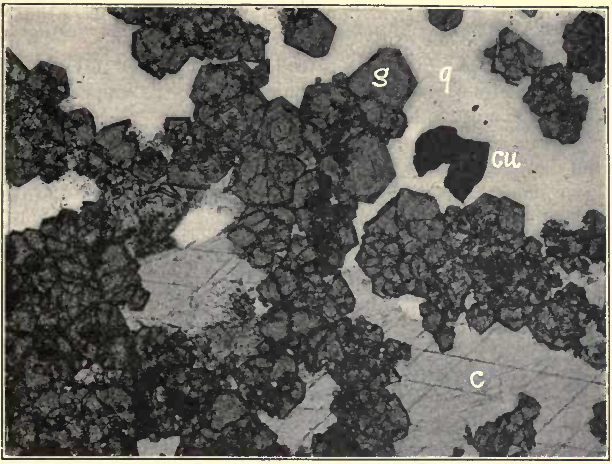

FIG. 250.—Thin section showing contact-metamorphic ore, Clifton, Arizona. *c*, Calcite; *g*, garnet; *q*, quartz; *cu*, chalcopyrite. Magnified 15 diameters.

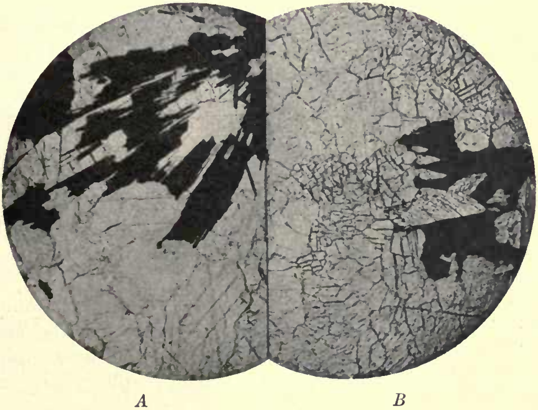

FIG. 251.—Thin sections showing contact-metamorphic ores at Holgol, Korea. *A*, Radiating crystals of ilvaite or hedenbergite (black) in granular limestone. Ilvaite incloses crystals of diopside. *B*, Chalcopyrite (black) in diopside. Magnified 30 diameters. *After B. Koto.*

in copper by oxidation, but in many occurrences the primary ore is rich enough to be utilized.

New Mexico.[1]—In the fourteen districts of New Mexico distinguished by contact-metamorphic deposits, the copper ores are by far the most common.

The most important of these deposits, economically, is that of the San Pedro mine, in the laccolithic mountain group of the same name. Beds of upper Carboniferous rocks over 700 feet thick have been metamorphosed by the underlying laccolith of granodiorite porphyry and by dikes extending upward from it. The lower 200 feet of shaly limestone is only partly altered, with local development of garnet and tremolite and a little chalcopyrite and pyrrhotite, but along a certain bed of purer limestone garnetization has taken place for half a mile, the thickness of this strongly metamorphosed stratum being about 50 feet. Bunches of chalcopyrite are irregularly distributed in it. Within this zone beds of pure crystalline limestone adjoin wholly garnetized beds. In places the rock consists of a mixture of garnet and coarsely crystalline limestone. On the dip the gently inclined ore beds have been followed for 300 feet. The upper beds of the series consist mainly of somewhat metamorphosed and baked shale and sandstone.

Clifton, Arizona.—In Arizona almost all the contact-metamorphic deposits yield copper as the principal metal. Near Clifton[2] a stock of granite porphyry and quartz monzonite porphyry breaks across pre-Cambrian granite, a Paleozoic series about 1,000 feet thick, and Cretaceous sediments about 400 feet thick. The Paleozoic limestones and shales, as well as the Cretaceous sandstones, are contact metamorphosed. The ore deposits lie mainly near Morenci and Metcalf; at both places the beds are cut by an unusual number of dikes, which have exerted a specially strong contact-metamorphic action on the sediments.

The ore deposits form a complicated series very similar to those observed at Cananea, Mexico, at Ely, Nevada, and at Bingham, Utah. The oldest ores are contact-metamorphosed limestones; these, as well as the adjoining porphyry, are cut by

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[1] Lindgren, Graton, and Gordon, *Prof. Paper* 68, U. S. Geol. Survey, 1910.

[2] W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905.

L. E. Reber, Jr., The mineralization at Clifton-Morenci, *Econ. Geol.*, vol. 11, 1916, pp. 528–573.

a series of pyritic veins, poor in copper, which in the sericitized porphyry spread out into disseminations of pyrite. Widespread oxidation has altered all the deposits and enriched them; well-defined chalcocite zones (p. 837) have formed by replacement of the pyrite by descending cupric sulphate solutions, and the present importance of the district is due wholly to the exploitation of these chalcocite ores, which contain from 2 to 4 per cent. of copper.

The primary contact-metamorphic deposits lie in limestone and form irregular bunches or tabular deposits parallel to dikes or following the stratification. Wherever the character is not masked by oxidation these primary ores consist of garnet, epidote, diopside, calcite, chalcopyrite, pyrite, magnetite, and zinc blende, occasionally also molybdenite. In the earlier days of the district, from 1875 to 1900, these oxidized ore-bodies were mined; they were easily reduced and comparatively rich in copper, containing mainly malachite, azurite, and limonite. The celebrated Longfellow ore-body formed a funnel-shaped mass in Ordovician limestone, between two porphyry dikes.

Farther west, in the Manganese Blue and Detroit mines, were several tabular ore-bodies, following the stratification in the Ordovician, Devonian, and Carboniferous limestones; these also owed their richness to several porphyry dikes, a few hundred feet from the main contact. Along the main contact were many irregular bunches of oxidized contact-metamorphic ores. At Metcalf the Shannon Mountain contained several similar ore-bodies (Fig. 246), lying in an isolated mass of Paleozoic limestones extensively cut by porphyry dikes.

Bisbee, Arizona.[1]—At Bisbee, Arizona, pre-Cambrian rocks are overlain by about 5,000 feet of Paleozoic limestones. After their deposition they were deformed by folding and faulting and were cut by intrusions of granitic porphyry of probably Jurassic Age, which is intimately connected with the origin of the copper deposits. The principal mass, of which the most prominent point is Sacramento Hill, close to Bisbee, is about

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[1] F. L. Ransome, *Prof. Paper* 21, U. S. Geol. Survey, 1904.

W. L. Tovote, *Min. and Sci. Press*, Feb. 4, 1911.

Arthur Notman, The Copper Queen mine and works, *Trans.*, Inst. Min. and Met. (London), vol. 22, 1913, pp. 550–562.

Y. S. Bonillas, J. B. Tenney and L. Feuchere, Geology of the Warren mining district, *Trans.*, Am. Inst. Min. Eng., vol. 55, 1917, pp. 284–355.

1 mile long and 1½ miles wide. After a period of erosion a transgression of the Cretaceous sea deposited a thick series of beds on the older series (Fig. 252).

The copper deposits lie in the limestones, surrounding crescent-like the east side of the intrusive mass, and appear as irregular and ill-defined or rudely tabular masses, which on the whole follow the dip of the stratification. They are almost entirely oxidized, even down to depths of 1,400 feet below the surface. The oxidized ore, containing copper carbonates, cuprite, and sometimes also chalcocite, with much limonite, passes gradually on its peripheries into "ledge matter" or limonitic clays, which in turn grade into altered and contact-metamorphosed limestones. The contact metamorphism is unusually inconspicuous, but the limestone surrounding the intrusive mass contains fine-grained tremo-

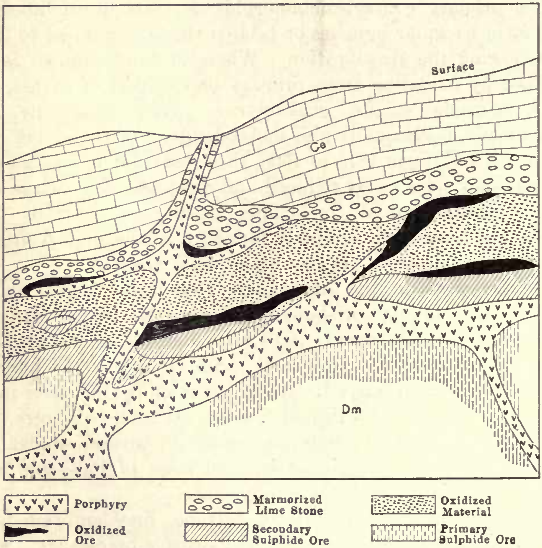

Fig. 252.—Generalized vertical section showing relations of primary ore, secondary sulphides and oxidized ore at Bisbee, Arizona. Ce, Escabrosa limestone; Dm, Martin limestone. *After J. Douglas and A. Notman.*

lite, diopside, garnet, vesuvianite, and quartz, associated with magnetite, pyrite, bornite, and a little chalcopyrite and galena and sphalerite. In the Oliver shaft, close to the contact, on the 1,200-foot level, the limestone is more intensely metamorphosed than elsewhere and contains bodies of pyrite, mixed with chalcopyrite and bornite.

The porphyry dips underneath the limestones and the contact is thus found farther west in the mines than on the surface. The limestones are also cut by a considerable number of dikes. A heavy mantle of pyritic quartzose ore, with some chalcopyrite and chalcocite, surrounds the porphyry, pitching parallel to the contact underneath the limestone of the surface.

The deep oxidization of the ores was practically completed before the Cretaceous and under conditions of water level different from those of today. The sericitized and chloritized porphrry contains some large bodies of chalcocite ore. Here, the mineral replaces pyrite.

The annual ore production of the Bisbee (Warren) district is now about 2,000,000 tons of 5 to 6 per cent., ore yielding, in 1916, about 200,000,000 pounds of copper with some lead, gold and silver; the total metallic value was $52,000,000. The principal production is derived from the Copper Queen and Calumet & Arizona mines.

Silver Bell, Arizona.—At Silver Bell,[1] southwest of Tucson, extensive primary chalcopyrite deposits have been worked. The mines are near the summit of one of the numerous desert ranges of that region; the ores were smelted, without concentration, at the Sasco plant. Several small masses of Paleozoic limestone are engulfed in a large mass of granite porphyry and along their contacts metamorphism is irregularly developed—in part by marmorization, in part by garnetization. The ore consists of chalcopyrite and light-brown garnet, said to be andradite, with a little magnetite, zinc blende, galena, and molybdenite. Much of the ore averaged 7 per cent. copper, but that smelted would average about 4 per cent. About 800 tons were mined per day in 1909. A trace of gold and 1 to 2 ounces of silver per ton are present. The oxidation is shallow, wholly fresh rock being encountered at the 200-foot level. The porphyry is locally silicified, but otherwise not greatly altered, except for some

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[1] C. F. Tolman, *Min. and Sci. Press*, Nov. 27, 1909.
C. A. Stewart, *Trans.*, Am. Inst. Min. Eng., vol. 43, 1913, pp. 240–290.

disseminated pyrite and chalcopyrite. No extensive chalcocite zone has been found in the porphyry.

Cananea, Mexico.—The mines at Cananea lie a short distance south of the Arizona-Sonora boundary line, in one of the short ranges that rise out of the gently sloping desert plains. Since 1900 these deposits have yielded large quantities of copper from ores enriched by oxidation and development of secondary chalcocite. The district was described by S. F. Emmons.[1]

The deposits show some similarity to those of Clifton, Arizona, but the geological history is much more complicated. Three successive irruptions, termed diorite porphyry, granodiorite, and quartz porphyry, have caused contact metamorphism in relatively small areas of Paleozoic limestone. Among the primary minerals are chalcopyrite, bornite, zinc blende, magnetite, and specularite; the limestones are garnetized, marmorized, and silicified.

A second epoch of mineralization by aqueous solutions resulted in veins and disseminations of pyrite, chalcopyrite, and quartz. Both classes have been enriched by oxidizing solutions.

Bingham, Utah.—The Bingham district, near Salt Lake, Utah, is now most widely known by the extensive mining operations of the Utah Copper Co. in chalcocitized porphyry (p. 866). According to J. M. Boutwell,[2] laccoliths and stocks of monzonite, as well as sills and dikes of diorite porphyry, invade the Carboniferous quartzite and limestone. Some of the ore-bodies are irregular replacements of contact-metamorphic type, others are later quartz veins containing lead, copper and zinc; both are altered and enriched by descending waters. Near the contacts the limestone is extensively marmorized and replaced by pyrite and chalcopyrite with local retention of its bedded structure, but the development of garnet and other silicates is unusually scant.

Ketchikan, Alaska.—Several contact-metamorphic copper deposits in southeastern Alaska are described by F. E. and C. W. Wright.[3] Those of Copper Mountain, Prince of Wales Island, present an unusually excellent illustration of deposits occurring at intervals along the contacts of an isolated granite intrusion.

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[1] S. F. Emmons, *Econ. Geol.*, vol. 4, 1910, pp. 312–356.

[2] J. M. Boutwell, *Prof. Paper* 38, U. S. Geol. Survey, 1905.

[3] *Econ. Geol.*, vol. 3, 1908, pp. 410–417. *Bull.* 347, U. S. Geol. Survey, 1908.

On the Kasaan Peninsula are several magnetite-chalcopyrite deposits, also containing pyrrhotite and pyrite, in a gangue of amphibole, epidote, orthoclase, garnet, and calcite. Wright believes that the ores were formed after the consolidation of the last intrusions of syenite. At both places shear zones and vein deposits containing copper accompany the contact deposits.

Zinc and Lead Deposits

Almost all contact-metamorphic sulphide deposits contain some zinc blende, and often also a little galena, but only a few deposits are known in which these metals constitute the principal value of the ore. Where they occur the amphiboles and epidote appear to be more prominent than garnet.

One of the best examples is furnished by the Magdalena mines,[1] in New Mexico, which in the oxidized zone, 200 to 300 feet deep, were worked for their lead, silver, and zinc. In depth large bodies of zinc blende were found, together with a little galena and chalcopyrite. According to Gordon the Magdalena Range consists of faulted blocks of Paleozoic (Mississippian and Pennsylvanian) limestone, resting on pre-Cambrian crystalline rocks. The limestones are cut by dikes of granite porphyry, which are exposed near the Graphic mine and which are believed to have caused the mineralization. In the limestones, which dip westward, toward a hidden contact with the granite porphyry, mineralization has taken place at five horizons, of which only one, just below the "Silver Pipe" limestone, is of great importance. The ore-bodies are roughly lenticular and may be as much as 40 feet in thickness. They occur at irregular intervals along the bedding planes, the principal bodies lying apparently at the crests of low arches transverse to the strike of the beds. Besides the sulphides mentioned they contain magnetite and specularite, with much epidote, pyroxene, and tremolite, but little if any garnet. The distance along the dip of the strata to the intrusive rock is probably not less than 2,000 feet.

Knopf[2] describes lead deposits from Darwin, Inyo County, California, which present an interesting succession ranging from contact-metamorphic types to fissure veins. Over large areas the Carboniferous calcareous rocks are altered to wollastonite,

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[1] C. H. Gordon, *Prof. Paper* 68, U. S. Geol. Survey, 1910, pp. 241–258.

[2] Adolph Knopf, *Bull.* 580, U. S. Geol. Survey, 1915, pp. 1–18.

diopside and grossularite rocks, with perfect preservation of structure. An enormous quantity of material, chiefly silica has been added during metamorphism. The contact-metamorphic ores lie between quartz-diorite and lime silicate rocks. The minerals are galena, andradite, calcite and fluorite. Orthoclase and apatite are present in other, similar deposits. Numerous veins of galena and fluorite, break through lime silicate rock in other parts of the district. The contact-metamorphic deposits are considered by Knopf to be later than the general metamorphism, but the argument is not wholly convincing.

Contact-metamorphic deposits have been described from eastern Mexico, where the thick Cretaceous limestone is broken through by many small intrusives. Most of these are copper deposits but sometimes they contain lead as at La Sirena Mine,[1] near Zimapan, Hidalgo, where dikes and masses of monzonitic character intruded into the pure limestone have caused mineralization along the contacts. The ore formation began by development of quartz, albite, danburite ($CaB_2Si_2O_8$), apatite, garnet, actinolite and fluorite in the order given. Then followed deposition of arsenopyrite and pyrrhotite, and lastly galena, zinc blende and jamesonite.

Finally there is to be mentioned the great Santa Eulalia[2] lead deposits near Chihuahua. These wonderful deposits which have yielded during the last two hundred years, lead and silver to the estimated value of $300,000,000 to $500,000,000, carry mainly oxidized ores and their geological affiliations have not been known until recently. They form tabular or irregular bodies or pipes in Cretaceous limestone and have been opened to a depth of 2,000 feet. These replacement ores are connected with N.-S. fissures. One type contains pyrite (replaced by pyrrhotite in depth), galena and zinc blende. The other type appears closely allied to the contact-metamorphic ores, and forms irregular masses. They are described as argentiferous pyritic ores and contain pyrrhotite with a little pyrite, galena, sphalerite, arsenopyrite, associated with iron manganese silicates such as ilvaite, knebelite, hedenbergite, fayalite and chlorite. The presence of intrusive bodies in depth is inferred, but so far none have been disclosed.

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[1] W. Lindgren and W. L. Whitehead, *Econ. Geol.*, vol. 9, 1914, pp 435–462.

[2] Basil Prescott, *Trans.*, Am. Inst. Min. Eng., vol. 51, 1916, pp. 57–99.

Gold Deposits

Gold is present in traces in almost all sulphide deposits of the contact-metamorphic type, and a few ounces of silver to the ton is likewise not unusual, but it is rare to find in such deposits ores which are valuable chiefly on account of their precious metals.

At the Santa Fe mine,[1] Chiapas, Mexico, limestone close to an intrusive contact is changed to garnet and wollastonite; this rock contains in disseminated form, or concentrated in ore channels, auriferous and argentiferous bornite. The gold is in part free. The average ore consists of 90 per cent. garnet and 10 per cent. quartz (and chalcedony) carrying from 3 to 4 per cent. copper, with 6 to 8 ounces in silver and $6 to $20 in gold per ton.

One of the best examples of a gold-bearing contact-metamorphic deposit is that of the Cable mine, Montana, described by W. H. Emmons.[2] The ores are included in a long, thin block of limestone, in contact on both sides with quartz monzonite. The principal minerals are calcite, quartz, pyrrhotite, pyrite, magnetite, and chalcopyrite, with actinolite garnet, and green mica. The gold, in part coarse, is disseminated in calcite, quartz, and sulphides. This deposit has yielded about $3,000,000.

Platinum is rarely found. One occurrence in Sumatra is mentioned by L. Hundeshagen;[3] the metal occurs in wollastonite.

Gold-Arsenopyrite Type.—The best example of this rare type, lately described by C. Camsell,[4] is the deposit worked by the Nickel Plate mine, British Columbia. Gently folded Carboniferous limestones associated with shale, quartzite, and volcanic tuffs are intruded by sheets of gabbro and diorite. Along the contacts of these sheets, and particularly of their apophyses, the impure limestones are converted into contact-metamorphic minerals with arsenopyrite. The commercial ore-bodies, which have yielded several million dollars in gold, are tabular and follow the dipping contacts of the basic rock, which are not exactly parallel

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[1] E. T. McCarty, *Trans.*, Inst. Min. and Met. (London), vol. 4, 1895–96, pp. 169–189.

Also H. L. Collins, *idem*, February, 1900, and *Trans.*, Am. Inst. Min. Eng., vol. 31, 1901, p. 446.

[2] *Prof. Paper* 78, U. S. Geol. Survey, 1913, pp. 221–231.

[3] L. Hundeshagen, *Trans.*, Inst. Min. and Met. (London), 1904, pp. 1–3.

[4] Geology and ore deposits of the Hedley district, B. C., *Mem.* 2, Canada Dept. Mines, Geol. Survey Branch, 1910.

with the inclination of the strata. The outside of the ore-body is irregular and gradually fading, reaching somewhat farther away from the contact in some beds than in others. The principal ore-body has been followed 350 feet along the dip and has a width parallel to the contact of 125 feet.

The ore minerals are, named in order of quantity, arsenopyrite, pyrrhotite, chalcopyrite, pyrite, zinc blende, tetradymite (Bi_2Te_3) and molybdenite. The depth of oxidation is slight, but in the upper levels free gold occurred associated with tetradymite, while at the greater depth now attained it seems to be intimately bound up with the arsenopyrite and is not amenable to amalgamation. The pure arsenopyrite may contain as much as 12 ounces of gold per ton. The gold tenor varies from $6 to $14 per ton, but beyond the ore-body minor quantities of gold are widely disseminated in the contact-metamorphic rocks. There is very little silver; traces of platinum (as sperrylite?) and nickel are present. The gangue minerals are andradite, pyroxene, epidote, calcite, and axinite, and the sulphides are closely intergrown with them, pointing to contemporaneous deposition; the chalcopyrite is somewhat later than the arsenopyrite. Quartz is distinctly later than the other gangue minerals. Orthoclase appears in gabbro dikes replacing pyroxene by a process of endomorphic contact metamorphism. Although the rocks are faulted and fissured by post-intrusive stresses, these fractures contain practically no valuable ores.

The only similar deposit described in the literature is that of Reichenstein, Silesia, the auriferous leucopyrite and arsenopyrite of which have been worked on a small scale, probably since the thirteenth century. According to C. Wienecke[1] and A. Bergeat[2] the ore-producing intrusive is probably a neighboring granite, and the altered rock a dolomitic limestone.

Telluride Type.—Contact-metamorphic deposits carrying telluride ores are rare. W. H. Weed[3] describes such an occurrence at the Dolcoath mine at Elkhorn, Montana, where auriferous tetradymite is found in a 15- to 18-inch bed of garnet, diopside, and calcite. Weed[4] also mentions a deposit at Bannock,

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[1] C. Wienecke, *Zeitschr. prakt. Geol.*, 1907, p. 273.

[2] Die Erzlagerstätten, vol. 2, 1906, p. 1137.

[3] W. H. Weed and J. Barrell, Elkhorn mining district, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, p. 506.

[4] *Trans.*, Am. Inst. Min. Eng., vol. 33, 1903, p. 732.

Montana, where tellurides occur with specularite, garnet, pyrite, and free gold at a contact between diorite and limestone.

The occurrence of altaite in the Cable mine, Montana, is mentioned by W. J. Sharwood.[1]

It seems well established, then, that tellurides may crystallize at high temperatures. They are not known as products of igneous consolidation.

Cassiterite Deposits

Contact-metamorphic deposits with the assemblage of fluorine and boron minerals characteristic of cassiterite veins are rare. The tin-bearing magmas, which are generally normal granites, seem to retain the tin and associated substances until a later stage, after consolidation of the rock.

Cassiterite occurs, in part in connection with fissures, in the contact-metamorphic deposits of Pitkäranta, in Finland; of Schwarzenberg and Berggiesshübel, in Saxony; and of Campiglia Marittima, in Tuscany. Other examples of more typical character have been noted at Dartmoor, in Devonshire, England,[2] and lately by A. Knopf[3] on Lost River, Seward Peninsula, Alaska. At the latter place the granite has produced a narrow but long contact zone of lime-silicate rocks rich in tourmaline, axinite, ludwigite, hulsite and paigeite (both ferromagnesian stannoborates), vesuvianite, fluorite, scapolite, chondrodite, galena, sphalerite, arsenopyrite, pyrrhotite, scheelite, magnetite, pyroxene, and cassiterite. In the banded limestone the argillaceous layers are converted to tourmaline, with tremolite and vesuvianite, while the purer calcareous layers are marmorized. The orbicular structure of the contact minerals of this district has been mentioned on p. 711. The deposit is said to be of little economic importance.

At this interesting locality cassiterite also occurs in tourmalinized granitic masses or dikes, in quartz veins cutting granite and developing greisen, in quartz porphyry dikes, and in quartz stringers cutting limestone and slate.

The dikes of quartz porphyry, which pierce the limestone, contain cassiterite, pyrite, arsenopyrite, wolframite, and fluorite,

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[1] *Econ. Geol.*, vol. 6, 1911, pp. 22–36.

[2] K. Busz, *Neues Jahrb.*, Beil. Bd. 13, 1899, p. 100.

[3] *Bull.* 358, U. S. Geol. Survey, 1908.

with mica and topaz. The adjoining limestone is reticulated by veins which carry cassiterite, and around these veinlets hornblende, vesuvianite, fluorite, plagioclase, mica, and topaz have formed.

In the so-called Dolcoath lode a narrow dike is transformed into danburite (borosilicate of calcium) and tourmaline, with some arsenopyrite and cassiterite.

In the offshoots from the main granite mass are found fluorite, cassiterite, muscovite, tourmaline, and topaz, the last two having crystallized after the feldspar and quartz.

These interesting observations clearly show the intimate connection and in fact the transition between contact-metamorphic deposits and veins.

TITANIUM DEPOSITS

Titanium minerals are rare in contact-metamorphic deposits. Rutile is occasionally present. Singewald[1] has lately described ores of this type from Cebolla, Colorado, which contain ilmenite and magnetite together with garnet, augite and calcite. From later information it appears that perowskite ($CaTiO_3$) and titanium garnet are also present.

SCHEELITE DEPOSITS

Scheelite ($CaWO_4$), a heavy white mineral of non-metallic luster, occurs in many contact-metamorphic deposits. Recently such scheelite deposits have been discovered in Mono, Inyo County,[2] California, which apparently are of economic importance. The scheelite is associated with garnet, calcite, hornblende and pyroxene. Metallic minerals are rare. In 1917 and 1918 several such deposits were discovered and worked in the Humboldt Range, Nevada, particularly near Mill City. The association here is calcite, scheelite, garnet, epidote and pyrite, and the deposit occurs in limestone close to granite and is intersected by a dike of aplite. Possibly many such deposits have been overlooked.

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[1] J. T. Singewald, Jr., *Econ. Geol.*, vol. 7, 1912, pp. 560–573.

[2] Adolph Knopf, *Bull.* 640, U. S. Geol. Survey, 1917, pp. 229–249.

Graphite[1]

Properties.—Graphite is a form of carbon crystallizing in the rhombohedral system; it is soft, is steel-gray to black, and has a greasy feel. Even in its purest form it contains a little volatile matter and ash, usually less than 1 per cent. Many varieties are impure, and for some purposes, like paint-making, material with as little as 35 per cent. graphitic carbon is utilized. Analyses quoted by Cirkel show that the commercial grades of graphite, even those considered as of high quality, contain several per cent. of volatile matter and may be high in ash. One analysis of graphite from Ceylon shows 5.20 per cent. volatile matter and 22.15 per cent. ash.

The question whether graphite really exists in some of the varieties of "graphitic slate" yielding "amorphous graphite" is debatable; the minuteness of the particles renders it difficult to determine whether they are crystalline or not. The term "graphitoid" has been proposed for such substances, but is not accepted by all authors. The best test for graphite is said to be its characteristic property of yielding "graphitic acid" ($C_{11}H_4O_5$) with strong oxidizing reagents such as nitric acid. The amorphous carbons do not respond to this test.

According to H. Moissan graphite begins to oxidize at 650° to 700° C. In texture graphite is flaky or scaly or, when in veins, is often fibrous perpendicular to the walls; these varieties are called "crystalline" in the trade. "Plumbago" and "black lead" are trade names for the mineral.

General Occurrence and Origin.—Graphite appears mainly in rocks which have suffered intense regional or igneous metamorphism. The literature on its occurrence and origin is very extensive and shows plainly that the mineral may have originated in several ways:

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[1] F. Cirkel, Graphite, its properties, occurrence, refining, and uses, Dept. Mines, Ontario, Canada, 1907, p. 307.

J. H. Pratt, G. O. Smith, E. S. Bastin and H. G. Ferguson in *Mineral Resources*, U. S. Geol. Survey, Annual issues, American Bibliog. in 1913, vol. 2, p. 245.

B. L. Miller, *Mineral Industry*, Annual issues.

E. Donath, Der Graphit, Leipzig and Vienna, 1904.

E. Weinschenk, Der Graphit, etc., Leipzig, 1904.

E. Weinschenk, Weitere Beobachtungen, etc., *Zeitschr. prakt. Geol.* 1903, pp. 16–24.

1. It may form an integral part of rock magmas and crystallize together with the rock. This possibility is indicated by its presence in meteorites, in the terrestrial iron of Ovifak, Greenland, in nepheline syenite,[1] and in pegmatites.[2] In some of the occurrences in pegmatite dikes the graphite has, however, clearly been absorbed from the surrounding crystalline limestone. This is the origin of one of the occurrences described by George Otis Smith, as well as of the graphite in a dike near Franklin Furnace, New Jersey, described by A. C. Spencer.[3] These deposits are rarely of economic importance.

2. Graphite forms by the recrystallization of carbonaceous matter in metamorphic sedimentary rocks resulting from sandstone, shale, limestone, or coal. This transformation is well established and can evidently be effected under conditions of intense regional or igneous metamorphism, but it probably requires a relatively high degree of heat, perhaps well above 200° C. The development of graphite in the zone of contact metamorphism is assumed by some authors, like E. Weinschenk, to mean that the carbon has been supplied by emanations from the magma. Weinschenk also applies this theory to its occurrence in many areas of regional metamorphism, but this view is probably not justified.

In studying the contact-metamorphic graphite from Ticonderoga, New York, E. S. Bastin showed by experiments that the contemporaneous quartz crystals had not been exposed to a temperature of 575° C. While a very high temperature is necessary for the manufacture of artificial graphite, the transformation can evidently be effected in nature at a much lower degree of heat.

3. Lastly, graphite occurs in veins, sometimes 2 or 3 feet wide, having the appearance of resulting from the filling of open fissures, and in this form the mineral usually possesses a marked transverse fibrous structure. Such veins are found in igneous rocks like pegmatites and granites, and also in the surrounding metamorphosed sediments. Fine examples are seen in the graphite regions of New York, Canada, and Ceylon.

The origin of this type is less easy to explain. As the veins are usually found near intrusive contacts where high heat pre-

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[1] T. H. Holland, *Mem.*, Geol. Survey India, vol. 30, 1901, p. 201.

[2] G. O. Smith, *Bull.* 285, U. S. Geol. Survey, 1906, pp. 480–483.

[3] *Geologic Folio* 161, U. S. Geol. Survey, 1908.

vailed, it may be conjectured that they were formed by deposition from gaseous carbon compounds, such as carbon monoxide or cyanogen compounds, perhaps with metals; in some of these graphites the ash contains much iron. The prevailing opinion is that the carbon is derived from surrounding sediments and was deposited shortly after the intrusion, but E. Weinschenk[1] and others consider it as originating from exhalations of igneous origin. The Ceylon veins, described by the same author, contain, in addition to graphite, quartz, rutile, orthoclase, apatite, pyroxene, and pyrite. Calcite is contemporaneous and intergrown with the graphite. Finally there are, both here and at other places described by Weinschenk, kaolin and the corresponding iron compound, nontronite, and these occurrences are held to support the theory of igneous derivation. This view is assuredly not justified, as the possibility that such highly hydrated compounds can be formed by igneous exhalations is decidedly remote (p. 328). Types 2 and 3 form many valuable graphite deposits.

Occurrences.—Deposits of graphite have been found in a number of States in the Union, but few are of economic importance; many of them are graphite slates or clays which are utilized as pigments or as lubricants.

A part of the domestic supply of "crystalline" graphite is obtained from New York; the mines are located in Essex, Warren, Washington, and Saratoga counties, in the Adirondack region,[2] and the largest mine, that of the American Graphite Company, has been worked for 30 years. The rocks are pre-Cambrian crystalline schists of sedimentary origin. The principal bed worked is a dark silver-gray quartz-graphite schist and is said to average about 6 per cent. graphitic carbon. Elongated quartz grains, muscovite, apatite, pyrite, and graphite, the latter in thin and ragged flakes, averaging about 1 millimeter in length, are the constituents. Two beds are known, one about 8 feet thick, the other from 3 to 20 feet. Excavations have extended for 2,000 feet along the gentle dip of the thicker bed, the greatest

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[1] *Abhandl.* Bayer. Akad. d. Wissensch., vol. 21, 1901, pp. 218–335.

[2] E. S. Bastin, *Econ. Geol.*, vol. 5, 1910, pp. 134–157.

D. H. Newland, New York State Museum, *Bulls.*, 1905 to 1916. Annual reports of the graphite industry.

Ida H. Ogilvie, *Bull.* 96, *idem.* (Geological map.)

H. L. Alling, The Adirondack Graphite Deposits, *Bull.* 199, New York State Museum, 1918.

depth below the surface being 250 feet. The associated rocks are garnetiferous gneisses and limestones of the Grenville series. The sediments are metamorphosed by intrusion and injection of granite and gabbro of Laurentian and possibly Algoman age.

Three miles northwest of Ticonderoga, in the same region, coarse graphite plates are irregularly distributed throughout the contact zone between pegmatite and pegmatitic granite and the schists and limestones which these rocks intrude. Contact-metamorphic minerals, like scapolite, pyroxene, and vesuvianite, occur in this zone. The graphite also forms veins, 1 to 2 inches wide, which cut across both the schist and granite. The deposits at this locality have been worked for a number of years.

In the last years the production of flake graphite from a belt of Paleozoic mica schist in Clay county, Alabama, and adjacent region has acquired considerable importance. The ore contains about 3 per cent. graphite.

A deposit containing graphite in veins similar to those of Ceylon has recently been found near Dillon, Montana.[1] The veins occur along a contact zone of granites and pegmatites, intrusive in pre-Cambrian schists and calcareous rocks which have been contact-metamorphosed.

At several places in New Mexico[2] intrusions of basic igneous rocks have altered the coal-beds of the Tertiary or Cretaceous formations. At Madrid the coal was converted to anthracite. Near Raton the intrusions have turned the coal into a coke-like material, but at one place 7 miles southwest of Raton a number of sills produced exceptionally intense metamorphism, converting the coal to graphite. Graphite also occurs in irregular masses in the diabase and has a more or less columnar texture normal to the faces of the igneous rock.

Similar conditions produced the important deposit of amorphous graphite of Santa Maria, in central Sonora, Mexico. According to F. L. Hess[3] several coal-beds, attaining a maximum thickness of 24 feet, have been subjected to contact metamorphism and folding by intruding granite and are converted into amorphous graphite. The main vein averages 86 per cent.

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[1] A. N. Winchell, *Econ. Geol.*, vol. 6, 1911, p. 218. E. S. Bastin, *Econ. Geol.*, vol. 7, 1912, p. 435.

[2] W. T. Lee, *Mineral Resources*, U. S. Geol. Survey, pt. 2, 1908, p. 733.

[3] *Idem*, p. 734.

graphitic carbon and furnishes a good material for the manufacture of lead pencils.

The graphite deposits of Ceylon[1] are among the most productive in the world, yielding about 38,000 short tons a year of high grade product. The mineral is said to occur as veins, varying in width from 12 to 22 centimeters. The mines are from 100 to 500 feet deep. The rough material often contains up to 50 per cent. impurities and is hand picked and sorted.

According to Bastin the veins are found in a fine-grained acidic or basic gneiss to which he applies the name granulite. The rock contains quartz, feldspar, garnet, pyroxene, biotite, etc. Some crystalline limestone is also present. The gneisses are intruded by granites and pegmatites. In the last few years Madagascar is rivaling Ceylon. In 1917, 35,000 tons are said to have been produced, the quality of flake graphite being about the same as that of the domestic output.

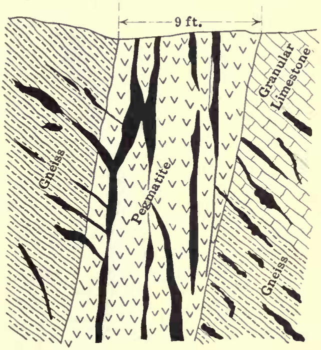

Fig. 253.—Vertical section of graphite veins, Buckingham Township, Quebec. *After A. Osann.*

The Siberian deposits, in the Batagol Mountains near Irkutsk, yield material of great purity, which formerly supplied the lead-pencil industry. L. Jaczewski,[2] describing the Alibert mines in this region, states that the graphite occurs in a nepheline syenite

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[1] J. Walther, *Zeitschr.* Deutsch. geol. Gesell., vol. 41, 1889, p. 359.
E. Weinschenk, *Op. cit.*
E. S. Bastin, *Econ. Geol.*, vol. 7, 1912, pp. 419–443 (with literature).

[2] *Neues Jahrb.*, 1901, 2, ref. p. 74. (Original in Russian.)

close to the contact of a schist that also contains graphite, the latter, as well as the inclusions in the igneous rock, being considered of organic origin. This conclusion is vigorously attacked by E. Weinschenk.[1]

The deposits at Passau, in Bavaria, comprise few veins; the graphite occurs in a crushed, schistose rock and Weinschenk regards the deposits as caused by volcanic emanations. The occurrences in Moravia are apparently similar.

The graphite deposits of Canada are contained chiefly in Buckingham and Grenville townships, Quebec, near Ottawa. The production in 1917 was about 3,700 long tons. These deposits, which have been described by A. Osann,[2] show particularly clear relations to contact-metamorphism. The rocks are largely gneiss, quartzite, and crystalline limestone of Grenville age cut by granite, pegmatite, and diorite. Graphite is widely distributed in fissure veins or lenticular masses in these intrusions or near their contacts, also as disseminated flakes in limestone or gneiss (Figs. 253 and 254). Associated minerals are apatite and scapolite, often appearing in the wall rocks of the veins,

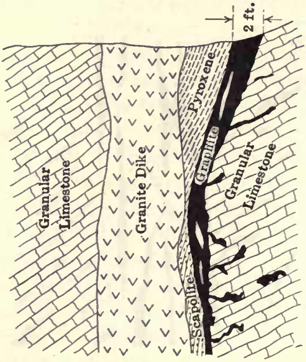

FIG. 254.—Vertical section of graphite vein in limestone, Grenville district, Quebec. *After A. Osann.*

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[1] E. Weinschenk, *Op. cit.*

E. Weinschenk, Zur Kenntniss der Graphitlagerstätten., *Abhandl.* Bayer. Akad. d. Wiss., vol. 19, 1899, pp. 511–563.

[2] *Ann. Rept.*, Canada Geol. Survey, 1899, pp. 660–820. See also Cirkel's report quoted above.

also biotite, titanite, wollastonite, and pyrite. The analogy of these deposits with the apatite veins is striking and the conclusion seems justified that they were developed by igneous emanations shortly after the close of the intrusive activity.

Production and Uses.—The production of natural graphite in the United States has varied considerably, owing to the large quantities of low-grade material used for paints and fertilizers. The output of flake graphite from New York State is about 1,500 tons per annum. Alabama, in 1917, produced 3,100 tons; the total domestic output of flake graphite in 1917 was 7,000 tons. Much larger is the production of artificial graphite now manufactured in electric furnaces at Niagara Falls at the rate of 5,000 tons per annum, from anthracite coal mixed with a small percentage of ash. In addition about 20,000 to 42,000 tons of graphite are imported from the highly productive mines in Ceylon, Mexico, Korea and Madagascar. Ceylon graphite sold in New York (1911) for 7 to 9 cents a pound, but during the war the price rose to 30 cents. Domestic No. 1 Flake brought 13 to 18 cents per pound in 1917. It should contain 90 per cent. graphitic carbon.

There is a great demand for graphite from many branches of industry. The inert and heat-resisting nature of the "crystalline" graphite makes it particularly valuable for crucibles, the fibrous Ceylon product being most suitable for this purpose.

Graphite is extensively used as a lubricant, with oil, and for this purpose the artificial mineral, which is "deflocculated," causing it to remain indefinitely in suspension in oil, is especially employed. Other uses are for pencils, foundry facings, polishing powder, paint, electrodes, and, strange to say, as an adulterant for fertilizers; it is claimed that it prevents absorption of moisture and caking.

The low-grade material from New York State is concentrated at the mines by crushing, washing on buddles or other appliances, and settling, but the details of the process have not been made public. Present practice in Clay County, Alabama, includes dry crushing, drying and water flotation.[1]

Garnet

Some varieties of garnet, especially almandite, are mined and used as abrasive material. In the State of New York there are

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[1] Irving Herr, *Eng. and Min. Jour.*, April 11, 1917.

several deposits of this kind.[1] The garnets occur in highly altered rocks of somewhat uncertain history but are probably the result of contact metamorphism.

DEPOSITS DUE TO IGNEOUS METASOMATISM NOT DISTINCTLY RELATED TO CONTACTS

General Features.—The deposits thus far described lie close to the well-defined contact of an intrusive rock with a sedimentary series. There are deposits, however, in which the mineral association points to the same mode of origin, but which are not clearly related to any given contact. This may result from a sloping or irregular contact of a large batholith, so that a point on the surface that is several miles from the contact horizontally may be only a few thousand feet from it in a vertical direction. General metamorphism, without special development of mineral deposits, appears to have been effected by such conditions at the northern end of the great batholith of Idaho between the Clearwater and St. Joe rivers.[2] During a long and deep immersion into the abyssal zone, metallic gases given off by magmas may have penetrated farther from the intrusion than they have near the surface. It is also possible that erosion may have cut away the metallizing dike or mass, so that its relation to the deposit is no longer apparent.

At any rate such ore-bodies are termed deposits due to igneous metamorphism, rather than contact-metamorphic deposits.

Ores of copper, zinc, lead, and iron are included in this class. Many representatives are found among the obscure deposits in the pre-Cambrian of Scandinavia.

Boundary District.—At Phoenix[3] and Greenwood, in British Columbia near the international boundary, are a number of ore-bodies which in the last decade have yielded about 125,000 tons of copper. The geology of the region is complex. A thick series of volcanic rocks (porphyrites), both clastic and massive, crystalline limestones, and argillites, all of upper Paleozoic age, is intruded by a granitic batholith of probable Jurassic age and smaller masses of syenite.

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[1] W. J. Miller, Garnet deposits of Warren County, N. Y., *Econ. Geol.*, vol. 7, 1912, pp. 493–501.

[2] F. C. Calkins and E. L. Jones, *Bull.* 530, U. S. Geol. Survey, 1913, pp. 75–86.

[3] O. E. LeRoy, *Mem.* 21, Canada Geol. Survey, 1912.

The large ore deposit of the Granby Company lies in a mineralized zone which represents a part of the limestone replaced by epidote, garnet, etc. The ore-bodies are lenses or large masses one of which is 2,500 feet long and 900 feet wide and has a maximum thickness of 125 feet. The dip becomes flat in depth and the ore ceases at a vertical depth of 675 feet. Fig. 255 represents the somewhat clearer condition at the adjoining Brooklyn Mine. The ore consists of chalcopyrite, pyrite, hematite, and magnetite, with andradite, actinolite, and epidote. Calcite and quartz fill the interstices between the lime-iron silicates. The ore as smelted contains from 1.2 to 1.6 per cent. copper with 0.04 ounce of gold and 0.3 ounce of silver per ton. The original limestone which appears in some remnants near the ore-body is comparatively pure and contains from 1 to 10 per cent. of silica and little or no iron. Magnetite, epidote, and garnet formed contemporaneously; somewhat later but partly overlapping came the development of chalcopyrite, pyrite, and hematite. The limestone is in large part converted to jasperoid, the alteration appearing to have taken place before the development of the ore.

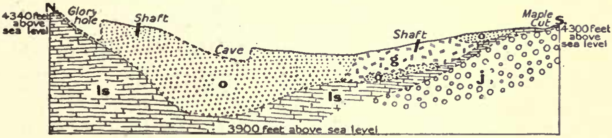

FIG. 255.—Generalized section of Brooklyn ore-body, Phoenix, B. C. *o*, ore-body; *ls*, limestone; *g*, gangue; *j*, jasperoid. Scale 400 feet = 1 inch.

No large bodies of igneous rocks appear in or near the deposits, and the nearest small outcrops of granodiorite are 1 to 2 miles away; one of these outcrops has been locally replaced by garnet, epidote, and actinolite. Deep drilling below the deposits failed to disclose intrusive rocks. It is held that the ores were formed by igneous emanations of iron, silica, and copper which traversed the limestone laterally from some unit of the intrusive series that is now eroded.

Ducktown, Tennessee.—The copper ores at Ducktown have been worked since about 1848 and still maintain an output of 8,000 tons of copper a year. In addition, about 700 tons of sulphuric acid is now obtained daily from these ores. The district, which lies in the mountainous area of the southern

Appalachians, has been the subject of repeated geologic investigation by C. Heinrich, J. F. Kemp, and W. H. Weed. Lately, W. H. Emmons and F. B. Laney[1] have examined the deposits.

According to Emmons and Laney the deposits are contained in a highly compressed metamorphosed and schistose series of arkose sediments of Cambrian age, consisting of poorly sorted conglomerates, grits, sandstone, and shale. Garnet and staurolite have developed abundantly in the rocks, the staurolite following certain horizons persistently. Thin lenses of limestone were contained in the series and are exposed in some places in the mines; they are now crystalline and contain layers of biotite and muscovite. Here and there are small ill-defined lenses of a highly metamorphic rock looking like a diorite-pegmatite and consisting of quartz, feldspar, hornblende, and garnet. These peculiar phases are now believed to be the result of strong metamorphism of the arkose sediments. Dikes of gabbro, later than the mineralization, are intruded in the sediments.

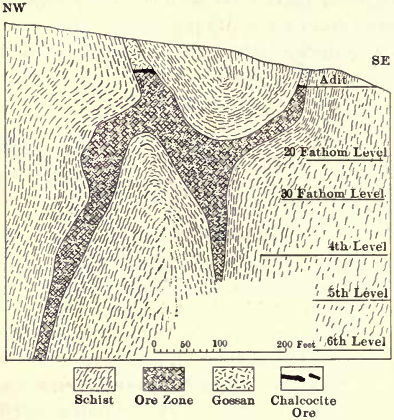

FIG. 256.—Cross-section of Mary mine, Ducktown, Tennessee. *After W. H. Emmons, U. S. Geol. Survey.*

The deposits are elongated, roughly tabular masses, some of them curved, lens-shaped, or folded, striking northeast and mostly dipping southeast (Fig. 256). The ore beds are parallel to the

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[1] Preliminary report in *Bull.* 470, U. S. Geol. Survey, 1911, pp. 151–172.

strike of the schists and average 60 feet in width. The primary ore is a coarsely crystalline mass of pyrrhotite, pyrite, chalcopyrite, zinc blende, specularite, magnetite, actinolite, calcite, tremolite, quartz, pyroxene, garnet, zoisite, chlorite, mica, graphite, titanite, and feldspars, all of practically contemporaneous crystallization.

Much of the ore is almost massive pyrrhotite and pyrite. Along the strike and dip the ore may grade into a lime silicate rock of gangue minerals and these in places grade into crystalline limestone. The contact between schist and ore is sharp or gradational within a few inches. The beds have been worked to a maximum depth of 1,000 feet. A thin but rich chalcocite zone due to enrichment by surface waters was found at a depth of 50 feet (p. 853), but below this the ores contain 1.5 to 3.0 per cent. copper, a small amount of silver, and a trace of gold. The ores from the Mary mine now average 2.5 per cent. It is held that the ores are formed by the replacement of thin limestone beds; all the abundant gangue minerals are in fact rich in lime. The replacement is believed to have been effected by igneous emanations, as a general association of minerals is typical of normal contact deposits. At the time of ore formation the rocks were at a high temperature and deeply buried, and it is thought probable that the emanations from some intrusion far below the surface, which had little effect on the schist, caused mineralization in the limestone beds. The mineralization fell within the epoch of dynamometamorphism; some deformation of the ore has taken place since its deposition.

Franklin Furnace, New Jersey.[1]—The great zinc-manganese deposits of northern New Jersey are of exceptional complexity and interest. Known since 1650 and actively worked since 1860, they now yield annually about 700,000 short tons of ore containing about 120,000 tons of zinc. The treatment of the crude ore by magnetic concentration yields franklinite, "half and half," and willemite; the first is used for the manufacture of zinc oxide for paints and leaves a manganiferous residue which goes to the blast furnace to make spiegeleisen; the second is also used for zinc white; and the third after further concentration yields a product of willemite from which a high-grade spelter (zinc) is made.

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[1] A. C. Spencer, H. B. Kümmel, J. E. Wolff, and Charles Palache, *Geologic* Folio 161, 1908.

See also review by C. K. Leith, *Econ. Geol.*, vol. 4, 1909, p. 265.

The two ore deposits of Mine Hill and Sterling Hill, 3 miles apart, are situated along a belt of pre-Cambrian crystalline limestone adjoined on the west by coarse gneisses of igneous origin. Cambrian limestone covers these rocks to the east and west. Both deposits form canoe-shaped beds in the pre-Cambrian limestone. The Mine Hill ore bed (Fig. 257) is closely adjoined along its west flank by the gneiss, the contact of which is parallel to the ore-body. The ore mass is thus a layer varying from 12 to 100 feet in thickness and, bent upon itself, forms a long trough or one-half of a canoe with sides of unequal height, the keel pitching north at a gentle angle.

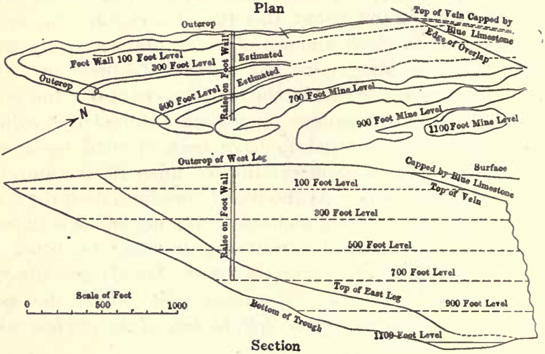

FIG. 257.—Plan of outcrop and levels and vertical section of Mine Hill ore-body, Franklin Furnace, New Jersey. *After A. C. Spencer, U. S. Geol. Survey.*

The mines are opened by a vertical shaft 965 feet deep and an incline 1,500 feet long. The ore forms transitions into the limestone and at Sterling Hill the limestone between the flanks also contains lean ore. Pegmatite dikes cut ore, limestone, and gneiss. The ore is a coarse aggregate of franklinite, 50 per cent.; willemite, 20 to 30 per cent.; zincite, 2 to 6 per cent.; and calcite, 3 to 11 per cent. Franklinite, $(Fe,Mn,Zn)O.(Fe,Mn)_2O_3$, contains about 42 per cent. iron, 15 per cent. manganese and 12 per cent. zinc; willemite, Zn_2SiO_4, 58 per cent. zinc; zincite, ZnO, 77 per cent. zinc. The four minerals mentioned are held to constitute the original ore. Besides, there are a great number of rarer minerals such as tephroite (Mn_2SiO_4), zinc pyroxene (schefferite), zinc amphibole, zinc spinel (gahnite), manganese garnet (polyadelphite), axinite (borosilicate of Al, Ca, Fe, Mn), apatite and scapolite (containing chlorine), rhodochrosite, fluorite, zinc blende, galena, arsenopyrite, chalcopyrite, and löllingite. Most of these minerals are regarded as products of secondary metamorphism due to the pegmatite dikes. Many veins cut the deposits, some of them containing the normal recrystallized ore minerals, others distinctly later with sulphides associated with calcite, albite, bornite, quartz, dolomite, etc.

In the older literature the deposits were considered of sedimentary origin. The question of genesis was reopened in 1889 by F. L. Nason, who admitted the possibility of igneous origin. Spencer believes that the original deposit was formed by the injection of magmatic emanations from the gneiss intrusions into the limestone. Participation in the general deep metamorphism which affected this region in pre-Cambrian time has further complicated the relations. It is certain that the texture of the ore and the universal rounding or corroding of the ore minerals point distinctly to igneous metasomatic action. The abundance of the spinel minerals is indicative of high temperature.

Metasomatic Magnetite Deposits of Sweden.[1]—Many of the earliest known and longest worked of the Swedish iron deposits

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[1] Hj. Sjögren, The genesis of our iron ores (Swedish), *Geol. För. Förhandl.*, vol. 28, 1906, pp. 314–344. With discussion by Törnebohm, Högbom, Holmquist, Bäckstrom, etc.

Hj. Sjögren, The geological relations of the Scandinavian iron ores, *Trans.*, Am. Inst. Min. Eng., vol. 38, 1908, pp. 766–835.

Hj. Sjögren, The question of the origin of the iron ores in the older pre-Cambrian series of Sweden, *Geol. För. Förhandl.*, vol. 30, 1908, pp. 115–155.

H. Johansson, The question of the origin of the middle-Swedish iron ores (Swedish), *Geol. För. Förhandl.*, vol. 28, 1906, pp. 516–538; vol. 29, 1907, pp. 143–186, 232–255; vol. 30, 1908, pp. 232–235.

Review, *Econ. Geol.*, vol. 5, 1910, pp. 494–498.

L. de Launay, L'origine et les caractères des gisements de fer scandinaves. *Ann. des Mines* (10), vol. 4, 1903, pp. 49–211.

See also a summary of recent literature by A. Bergeat in *Fortschritte der Mineralogie, etc.*, vol. 2, Jena, 1911, pp. 43–44.

Excellent descriptions of individual districts are found in the guide to the excursions of the Internat. Geol. Congress, Stockholm, 1910.

P. Geijer, Some problems in iron ore geology in Sweden and America, *Econ. Geol.*, vol. 10, 1915, pp. 209–239.

P. J. Holmquist, Swedish archaean structures and their meaning, *Bull.*, Geol. Inst. Upsala, vol. 15, 1916, pp. 125–148.

P. J. Holmquist, Structure and metamorphism of Swedish iron ores. (Swedish) *Geol. För. Förhandl.*, April, 1913, pp. 233–272.

form irregular masses or lenses in rocks of upper Archean age. They are either directly associated with crystalline limestone, or they occur near limestone but surrounded by masses of silicates like pyroxene, garnet, and epidote, to which the term "skarn" is usually applied. Though not as large as some of the more recently discovered deposits of certain or probable magmatic origin, the deposits have in the aggregate furnished much ore of exceptional purity and as yet are far from being exhausted. Until about ten years ago these deposits were considered by the Swedish geologists as of sedimentary origin, like bog iron ores, but subsequently metamorphosed. In modified form this opinion was expressed by de Launay in 1903. At present few observers hold to this view. There is, for instance, a strong

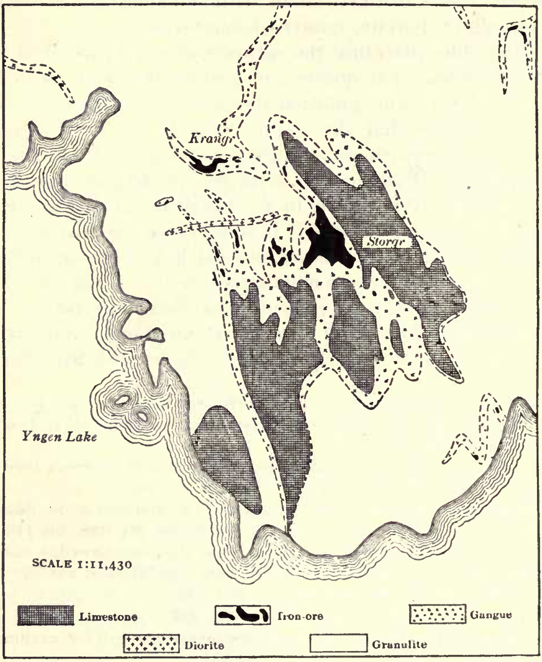

Fig. 258.—Plan of the Persberg mines, Sweden. *After Hj. Sjögren.*

similarity between the Swedish ores and those of the Banat province of Hungary, first described by von Cotta, and the latter are generally accepted as of contact-metamorphic origin. Striking and unmistakable also is their similarity to the metasomatic contact deposits of North America, many of which contain much magnetite and which at some places are worked for iron.

The Swedish deposits are, however, not so simply explained, for while in the districts mentioned the ores unquestionably adjoin igneous intrusions, the granitic rocks of Sweden are generally later than the deposits, which normally are contained in a peculiar fine-grained rock with gneissoid texture that is widely distributed in the iron region and that has been variously designated "hälleflinta," eurite, leptite, or granulite. These rocks, which form wide zones in the pre-Cambrian of Sweden and are locally associated or interbedded with amphibolites and smaller masses of more distinctly sedimentary quartz-mica slates and also with lenses of crystalline limestone or dolomite, are salic rocks, generally with at least 67 per cent. silica, and consist largely of albite, orthoclase, and quartz. Johansson has shown that they are in part potassic, in part sodic, and that intermediate composi-

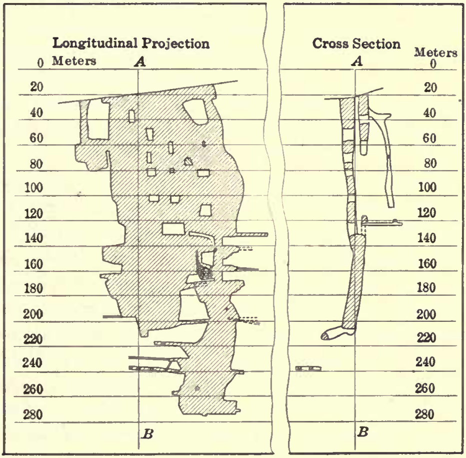

FIG. 259.—Vertical sections of the Kran mine, Persberg, Sweden. Shaded areas indicate stopes. *After Hj. Sjögren.*

tion is rare. He therefore interprets them as highly differentiated intrusives. The striped structure is interpreted by him as the result of a mechanical churning of the magma during differentiation. The most prevalent opinion is that these rocks are in part effusive, perhaps originally tuffaceous, and in part intrusive, and that the limestone and mica schist are of sedimentary origin.

The bodies of magnetite are in general associated with masses of crystalline limestone in this leptite formation, and it appears as if the mineralization were caused by the action of the granulite on the limestone. The ores form stock-like masses with greatest extension in a vertical direction and border directly against granulite, limestone, or "skarn." The bodies have been followed to a depth of about 1,000 feet; some of them cease distinctly at various depths but other stocks still continue below the greatest depth reached. Many of them, but not all, conform with the banding of the leptite (Figs. 258 and 259).

The "limestone ores" are more directly embedded in limestone, but here too skarn minerals may be present. In such an ore-body at Klackberg a narrow zone of dark-brown garnet was noted along the contact of limestone and ore and in the limestone itself was disseminated a dark-brown amphibole. The limestone ores often carry manganese and some of them constitute manganese deposits like that of Långbanshyttan, at which an unusually great number of rare minerals are found. Stretching and schistosity were superimposed upon the deposits in places and sometimes the direction of the stretching indicates the pitch of the ore-body. The magnetite is fine-grained; it contains in places a little specularite. Some deposits contain small quantities of pyrite, pyrrhotite, chalcopyrite, and arsenopyrite.

The composition of one of the Persberg ores is as follows:

Fe_3O_4	71.56	CaO	4.85
FeO	5.11	Al_2O_3	0.77
Fe	55.79	SiO	12.76
MnO	0.17	P_2O_5	0.005
MgO	4.18	S	0.031

Secondary changes have resulted in crushing along certain zones (skölar) and a great development of chlorite and other minerals of dynamometamorphic affiliations. Among the celebrated deposits of this type should be mentioned those of Persberg, Taberg (in Wermland), Nordmark, Norberg, and Dannemora.

The field relations indicate beyond doubt that the ores and skarn are metasomatic replacements of limestone or dolomite similar to contact-metamorphic deposits, probably effected by very hot solutions containing iron, manganese, silica, etc., derived from intrusive magmas.

Holmquist holds that the bedded, supracrustal leptites with accompanying sedimentary iron ores (p. 825) subsided into underlying granitic magma, which effected igneous metamorphism of the bedded ores and developed magnetite and lime silicates in the limestone. The later events included a regional metamorphism which affected the ores to some degree, and the final intrusions of granite, pegmatite and diabase which have exerted very slight influence on the deposits.

Magnetite Deposits in the United States.—Deposits of magnetite which are similar to the Swedish ores just described are found in the United States at few places. The Tilly Foster mine,[1] in New York State, contained a steep lenticular body of ore embedded in gneiss; the magnetite was associated with calcite, dolomite, chondrodite, enstatite, epidote, chlorite, garnet, and scant sulphides. The ore-body was followed to a depth of about 600 feet.

Another locality is in the Cranberry district, in North Carolina, described by A. Keith,[2] where low-grade ores have been mined and concentrated. The ore here occurs as a series of lenses of magnetite in a gangue of hornblende, pyroxene, and epidote; the lenses dip southwest at angles of 45°, about parallel to the schistosity of the surrounding gneiss. The ore is pure, with little phosphorus. It is not certain whether it represents replaced limestone.

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[1] F. R. Koeberlin, The Brewster iron-bearing district of New York, *Econ. Geol.*, vol. 4, 1909, pp. 713–754.

[2] *Bull.* 213, U. S. Geol. Survey, 1903, pp. 243–246; also in *Folio* 90, U. S. Geol. Survey.

CHAPTER XXVIII

MINERAL DEPOSITS OF THE PEGMATITE DIKES

INTRODUCTION

Each large intrusive mass is usually accompanied by a series of later dikes. These "complementary" dikes have, as a rule, a composition similar to that of the prevailing rock, but differ from it in showing an enrichment of certain constituents and a reduction of others. They are generally regarded as products of magmatic differentiation, forming residual parts of the dominant magma after its consolidation has begun. Some of them are basic, like kersantite, minette, or camptonite; others are acidic, like granite porphyry, aplite, or pegmatite.

Under the name of pegmatite are grouped the coarse granular dike rocks, often with well-developed idiomorphic texture, which accompany intrusive rocks, each group being characterized by pegmatites of special types.

Gabbros are sometimes accompanied by basic pegmatites of feldspar and pyroxene, and diorite by similar dikes of a basic feldspar and hornblende. The anorthosites are followed by pegmatitic dikes containing labradorite, hypersthene, and ilmenite; the nepheline syenites by pegmatites of soda feldspars, nephelite, sodalite, lepidomelane mica, ægirine, arfvedsonite, and minerals containing zirconium and titanium.

Most abundant are the granitic pegmatites, which consist mainly of coarsely crystallized orthoclase and quartz with muscovite; they often contain tourmaline, cassiterite, monazite, orthite, topaz, and a host of other rare minerals.

MINERALIZERS AND THE NATURE OF THEIR ACTION[1]

The processes of intrusion and crystallization bring about an increasing concentration of the volatile constituents of the rock magma, if no other avenue of escape is open to such substances.

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[1] In part after A. Harker, Natural history of igneous rocks, 1909, pp. 282–302.

W. O. Crosby and M. L. Fuller, Origin of pegmatite, *Am. Geologist*, vol. 19, 1897, pp. 147–180.

In subaerial eruptions they are given off into the atmosphere. These volatile substances, which of course formed an integral part of the original magma, consist of water and compounds of boron, fluorine, chlorine, phosphorus, sulphur, carbon, arsenic, tellurium, selenium and also other rarer elements. They exert a peculiarly favorable action upon the crystallization of magmas and minerals by decreasing their viscosity, lowering their freezing point, and furthering the development of minerals which otherwise do not crystallize from dry magmas. Harker says:

> The action is doubtless partly physical, partly chemical. The nature of the chemical effect, where the agent does not enter the crystallized product, is sometimes designated as a *catalytic* action, signifying a peculiar property possessed by certain bodies of inducing chemical changes in other bodies without themselves entering into the composition of the final product. In other instances the "mineralizer" forms part of the crystallized material.

French investigators from the days of Élie de Beaumont[1] have justly laid stress on the part played by mineralizers in magmatic differentiation and in the formation of mineral deposits. In the acidic rocks, which are known to contain fluorine and boron, the action of mineralizers is particularly clear, but they are doubtless present also in basic rocks, in which chlorine, phosphorus, and sulphur take the place of fluorine and boron. Some water is probably always present, although the tendency of some investigators is to minimize its importance.[2]

The presence of inclusions containing water in quartz crystals of acidic intrusive rocks shows plainly enough that the magma contained some water, as do also the transitions from pegmatite dikes to deep-seated ore-bearing veins.

The residual magma contains, besides these volatile mineralizers, the principal elements of the igneous rock crystallizing as quartz, feldspar, ferromagnesian minerals, and muscovite, and a number of rarer elements, such as tin, tungsten, zircon, tantalum, columbium, cerium, beryllium, molybdenum, lead, copper, lithium, and cæsium. These rare elements appear to have been carried along, in the process of differentiation by the mineralizers, which in many cases have also carried large quantities of iron differentiated from the main igneous body.

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[1] Note sur les émanations volcaniques et métallifères, *Bull.* Soc. géol. France (2), vol. 4, 1847, pp. 1249–1333.

[2] A. Brun, Recherches sur l'exhalaison volcanique, Geneva, 1911.

The mineralizing agents do not confine their action to the later stages of differentiation, but doubtless play a part in the crystallization of the main body of every magma. This is shown by the occurrence of molybdenite, pyrite, bismuthinite, zinc blende titanite, and zeolites in the druses of granitic rocks; among such are the occurrences of Striegau, described by A. Schwantke,[1] and those in the syenitic rocks in the vicinity of Kristiania, mentioned by Goldschmidt.[2] The granites of the Island of Elba contain in druses such minerals as albite, tourmaline, beryl, garnet, pyrite, arsenopyrite, cassiterite, and zeolites.[3]

Many of the silicate minerals, formed by the aid of mineralizers as the last stage of intrusive action, are of remarkably complex chemical nature. To many minerals of this class no formula can confidently be assigned. Other minerals, especially the sulphides, are characteristically of simple formula and composition.

A distinct paragenesis or succession of minerals is noted in many pegmatites. With successively lower temperatures new sets of minerals were formed and many of those stable at a higher degree of heat became subject to alteration as the temperature became lower. Thus in the Norwegian pegmatite dikes Brögger distinguishes three epochs of crystallization ending with the zeolites.

TEMPERATURE OF CONSOLIDATION

The residual magma, then, contains an increased quantity of mineralizers and their accompanying metals and has also a lower temperature than the original magma, in some cases doubtless lower than 500° C. It is injected into the earlier consolidated magmas and also into the encasing rocks; its fluidity and low melting-point are factors of great importance, allowing it to completely soak and penetrate schistose and fissile rocks encountered in its way. The pegmatites are essentially residual magmas, but they may become so admixed with water and dissolved gases that we may speak of them as in aqueo-igneous fusion at a temperature of 300° or 400° C. Even at this point with pressures over 200 atmospheres the critical point of water is exceeded.

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[1] Drusenmineralien des Striegauer Granits, Leipzig, 1890.

[2] V. M. Goldschmidt, Die Contactmetamorphose im Kristiania Gebiet, Kristiania, 1911.

[3] G. Vom Rath, Die Insel Elba, *Zeitschr.* Deutsch. geol. Gesell., vol. 22, 1870, p. 466.

OCCURRENCE AND GENERAL CHARACTER

The pegmatites form dikes, sheets, pipes, and irregular masses; where appearing as dikes or sheets no great regularity or extended continuation in depth can be counted upon, and this is important to consider in the exploitation of such bodies. Probably this irregularity is explained by the sudden and explosive action by which they make room for themselves and hold the cavities open until their substance is crystallized. Very different is this action from the slowly applied compressive stresses by which the fissures of most veins are opened.

The pegmatites are essentially coarsely crystalline rocks. Under some circumstances the dimensions of the crystals may be enormous. In the Ural Mountains a quarry was opened in a single orthoclase crystal; in India muscovite plates 10 feet in diameter have been found; at the Etta mine, in the Black Hills of South Dakota, spodumene occurs in crystals resembling tree trunks and as much as 42 feet in length; quartz crystals several feet in length are not uncommon. Often the minerals crystallize together, as feldspar and quartz in graphic granite, but in other pegmatites there is a distinct succession, with muscovite, for instance, at the walls and quartz and feldspar in the center, or with feldspar crystals along the walls and a central filling of quartz. The rarer minerals usually form the later generations and probably crystallized below 575° C., the crystallographic inversion point for quartz. The pegmatites are evidently not eutectics. They crystallized under the same general pressure and temperature as the magma itself. The rarer minerals are accessory, as a rule, for there are enormous masses of pegmatites which contain little but quartz and feldspar.[1]

In their present condition there is little evidence of water as a constituent of their magma, but facts already referred to force us to the belief that some water was present as well as some carbon dioxide.

Liquid inclusions in pegmatitic quartz from Branchville, Connecticut,[2] were found to consist of 98.33 per cent. CO_2, 1.67 per

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[1] E. S. Bastin, Origin of the pegmatites of Maine, *Jour. Geology*, vol. 18, 1910, pp. 297–320. *Bull.* 445, U. S. Geol. Survey, 1911.

W. T. Schaller, Southern California pegmatites, *Science*, April 1, 1910.

[2] G. W. Hawes and A. W. Wright, *Am. Jour. Sci.*, 3d ser., vol. 21, 1881, pp. 203 and 209.

cent. nitrogen, and traces of hydrogen sulphide, ammonia, fluorine, and possibly chlorine.

A marked contact-metamorphic action, sometimes stronger than that of the original magma, characterizes many pegmatites. Here, too, as in the case of normal igneous rocks, it is well to distinguish between the ordinary contact metamorphism without additions of material, and metasomatic contact metamorphism, in which substances contained in the pegmatite penetrate into the surrounding rock and replace some of its minerals. H. B. Patton[1] describes, for instance, a pegmatite dike in Colorado which is 10 feet wide and which contains but little tourmaline, but which strongly impregnates the surrounding rock with this mineral for a distance of 2 or 3 feet from the contact. It is, however, worthy of note that no sulphide impregnations analogous to the normal contact-metamorphic deposits have been found at the contacts of pegmatite and limestone. The quartz monzonites of the Western States, along whose contacts most of the deposits mentioned occur, are rarely accompanied by pegmatitic dikes.

The pegmatites often absorb material from their walls, and near them minerals otherwise foreign are likely to appear; andalusite, garnet, and staurolite are among these minerals.

TYPES OF PEGMATITES

Acidic Pegmatites.—The most common type consists of the granitic pegmatites, which always contain orthoclase, albite, and quartz, usually also microcline and muscovite. Among the accessory minerals magnetite, often in crystals, is perhaps most common. Other rarer minerals are tourmaline, topaz, fluorite, cassiterite, apatite, ilmenite, rutile, orthite, monazite, beryl, samarskite, spodumene, amblygonite, and many more. The typical mineralizers are boron and fluorine, together with a little phosphorus and sulphur. Lithium and the metals of the cerium and thorium groups are also characteristic. Among sulphides molybdenite and bismuthinite are the most common, but pyrite, arsenopyrite, pyrrhotite, chalcopyrite, bornite, and sphalerite are also found. In the south Norwegian granitic pegmatites lithium and tin are absent.[2]

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[1] *Bull.*, Geol. Soc. Am., vol. 10, 1898, p. 21.

[2] W. C. Brögger, Die Mineralien der südnorwegischen Granit-pegmatit Gänge, *Videnselskabets Skrifter*, Math-naturv. Klasse, Kristiania, 1906, No. 6, pp. 159.

A second group is formed by the syenitic pegmatites, rich in alkalies and especially in sodium. These contain soda orthoclase, ægirine, acmite, arfvedsonite, biotite, nephelite, sodalite, låvenite, and a number of rare titanium and zirconium minerals, also fluosilicates. There is little or no quartz. The characteristic mineralizers are fluorine and chlorine. Here, as elsewhere, the sulphides belong to a rather late stage of consolidation. Brögger[1] distinguishes in south Norwegian syenite pegmatites four phases of crystallization. In the first phase (of earliest development) he places feldspars, nephelite, sodalite, ægirine, lepidomelane, barkevikite, and magnetite, followed by fluorite, rosenbuschite, låvenite, and woehlerite (containing fluorine); by sodalite (containing chlorine); by helvite (containing sulphur); by löllingite (containing arsenic); and by homilite and melanocerite (containing boron). There is no tourmaline, topaz, or quartz.

The second phase consists in the filling of drusy cavities in part by destruction of the older minerals; these druse minerals consist of leucophane and fluorite, representing the fluorine group; of homilite and datolite, representing the boron group; and of the simple sulphides, such as molybdenite, sphalerite, and galena.

The third phase, at a considerably lower temperature but still probably above 100° C., comprises the zeolites, which are followed by a fourth phase of low-temperature carbonates and fluocarbonates.

Interesting pegmatite pipes in the riebeckite granite of Quincy, Massachusetts, have been described by Warren and Palache.[2] A zone of a coarse granitic aggregate of quartz, feldspar, riebeckite, and ægirine graduates into a central mass of almost pure massive quartz, sometimes containing molybdenite, sphalerite galena, and chalcopyrite and in its miarolitic cavities fluorite, octahedrite, ilmenite, and parisite, the last a fluocarbonate of the cerium metals.

Basic Pegmatites.—The basic pegmatites are less common. Boron and fluorine are not usually present, but phosphorus and chlorine, probably also sulphur, play important parts. Soda-lime feldspars, amphibole, pyroxene, quartz, apatite, rutile, and brown mica are the most abundant minerals.

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[1] W. C. Brögger, *Zeitschr. Kryst. Min.*, Bd. 16, 1900.

[2] *Bull.*, Geol. Soc. Am., vol. 21, 1910, p. 784.

ECONOMIC FEATURES OF PEGMATITE DIKES

The pegmatites, on one hand, contain many of the common minerals in exceptional size of grain and purity, and, on the other hand, they are a storehouse for a great number of the rarest minerals, many of which are not found elsewhere. These deposits are therefore of considerable economic importance and their valuable products are of manifold kind.

Feldspar and Quartz.—The orthoclase and quartz of granitic pegmatites are mined or quarried at numerous places, particularly in Maine, Connecticut, and Pennsylvania, in Norway, and in many other countries. The total value of the quartz and feldspar obtained from pegmatite dikes in the United States amounts annually to several hundred thousand dollars. The minerals are used for pottery and for many other industrial purposes; the quartz in particular is also utilized as an abrasive, in paints, and for the coating of tarred roofing. A minor quantity of quartz is also cut as a semi-precious stone under the names rock crystal, smoky quartz, rose quartz, and rutilated quartz.[1]

Mica.—White mica, more commonly known as muscovite, is also an important product of the granitic pegmatites. It occurs as irregularly disseminated bunches of foils, or "books," in pegmatite dikes, sometimes crystallizing along the walls. The mica-bearing pegmatites are worked in three belts in the North Carolina mountain region, where they break into the pre-Cambrian crystalline schists and gneisses.[2] The dikes, which also carry orthoclase, perthite, oligoclase, and quartz, are of varying thickness and persistence, at some places lenticular and

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[1] D. B. Sterrett, Gems and Precious Stones, *Mineral Resources,* U. S. Geol. Survey, pt. 2, 1910, pp. 963–975; also in later issues.

[2] D. B. Sterrett and W. T. Schaller, *Mineral Resources*, U. S. Geol. Survey, 1908–1917.

D. B. Sterrett, Mica deposits of South Dakota, *Bull.* 380, U. S. Geol, Survey, 1909, pp. 382–397.

D. B. Sterrett, Mica deposits of North Carolina, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 400–422.

J. A. Holmes, *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 6, 1899, pp. 691–707.

R. W. Ells, Mica deposits of Canada, Mineral deposits of Canada, 1904.

F. Cirkel, Mica, its occurrence, exploitation, and uses, Dept. of Interior, Mines Branch, Canada, 1905.

A. Osann, Notes on certain Archean rocks of the Ottawa valley, *Ann. Rept.*, Canada Geol. Survey, vol. 12, 1899.

following the schistosity, at others cutting across the country rock. Accessory minerals are biotite and several rare or gem minerals, among them beryl and its variety aquamarine. Some of the occurrences constitute transitions to quartz veins, which are assumed to have been formed by more distinctly aqueous solutions.

Muscovite is also mined in New Hampshire, Virginia, South Dakota, Alabama and Georgia.

Muscovite owes its usefulness to its transparency, elasticity, great resistance to heat and weathering, and applicability as a non-conductor of electricity. Its crystals are sometimes several feet in diameter, but this is exceptional, sheets of 1 foot in diameter being considered large. Smaller sheets a few inches square find ready use and are split into thin lamellæ and cut into proper shapes for stove doors and for various electrical insulating purposes. The scrap from the trimming is often ground and compressed into mica board or "micanite" for use in insulating. It is also used for the manufacture of wall papers and roofing materials. The quality of mica is best judged by the transparency of sheets about 2 millimeters thick; it is graded as "wine" or "rum" or smoky and spotted mica, the latter being undesirable for insulation. The price paid for sheet mica varies greatly according to the size of the sheets. The average

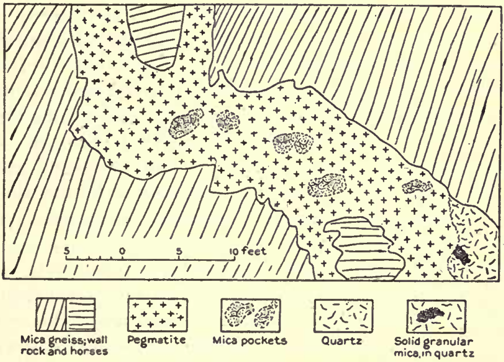

FIG. 260.—Vertical section across pegmatite dike, Thorn Mountain mine, N. C. *After D. B. Sterrett, U. S. Geol. Survey.*

price ranges up to $1 per pound for sheets 2 by 3 inches; larger sheets are worth several dollars per pound. In 1917 the domestic output was 1,217,000 pounds of sheet mica and 3,250 tons of scrap. Considerable quantities are imported, mainly from Canada and from India.

The Indian mica, which is mined on a large but primitive scale, is generally muscovite contained in pegmatite dikes cutting gneissoid rocks. The Canadian mica, of which much is also exported, is mainly a phlogopite or brown magnesium mica and is better adapted to electrical uses than the muscovite. It occurs with apatite, in dikes or veins of pyroxene in gneiss or limestone, the principal localities being north of Ottawa. Associated minerals are calcite, scapolite, titanite, various metallic sulphides, among which molybdenite is mentioned, and one or two zeolites. These peculiar deposits are undoubtedly analogous to the Norwegian apatite-scapolite veins; in part they are certainly derived from limestone by contact metamorphism. Many of these dikes have been worked to a depth of several hundred feet. The quantity of trimmed mica obtained from the rock mined is small, often less than 1 per cent. Occasionally plates 5 feet in diameter are found.

Oxide Ores.—Specularite, magnetite, and ilmenite are of common occurrence in pegmatites, but scarcely ever of economic importance. Cassiterite, or oxide of tin, is also very common in many granitic pegmatites. Cassiterite in many places forms an integral constituent of granites, having unquestionably been consolidated with the other magmatic minerals. Its most common occurrence is in quartz veins which were formed at high temperatures, as indicated by the mineral association, but which differ from the pegmatitic dikes and assuredly were formed at somewhat lower temperatures than the dikes. These veins, too, stand in closest areal connection with the acidic intrusive rocks.

Pegmatites containing cassiterite, with phosphates and lithium minerals, have been mined near Gaffney, South Carolina, and about 50 tons of tin have been obtained from the detrital deposits.[1] The average tenor is low, but the mineral is concentrated along certain lines in the dike not unlike a shoot in a metalliferous vein. Tin-bearing pegmatites occur also in the Black Hills of South Dakota, where attempts to mine them have showed

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[1] L. C. Graton, Gold and tin deposits of the southern Appalachians *Bull.* 293, U. S. Geol. Survey, 1906.

that they carried a very low percentage of the metal.[1] At Tinton, in the northern Black Hills, mining operations have been carried on and some cassiterite recovered. At the Etta or Harney Peak mine, in the southern Black Hills, the percentage of tin appears to be too small for successful recovery, but other minerals, particularly those of lithium, have been mined (p. 773).

In the New England district in New South Wales, as described by E. C. Andrews and L. A. Cotton,[2] there are pipes of greisen with transitions into pegmatite containing cassiterite associated with wolframite, molybdenite, bismuth, arsenopyrite, tourmaline, fluorspar, and beryl.

Remarkable rutile deposits have been discovered recently in Virginia, in Amherst and Nelson counties.[3] They are probably pegmatitic developments of probably pre-Cambrian gabbro magmas, which, in other parts of the world, are also characterized by the concentration of titanium and phosphorus.

It is apparently a case where it is difficult to draw the line between ordinary rock differentiation and pegmatization, but the features of the deposits clearly recall the latter process. The districts mentioned contain a predominant rock of quartz monzonite gneiss with an unusually large percentage of titanium and phosphorus. Besides there are dikes of gabbro still richer in titanium. The pegmatitic facies consist essentially of a bluish quartz with plagioclase, orthoclase, and pyroxene, the last converted into hornblende with much rutile and accessory apatite and ilmenite. The rock has an even granular texture and contains as much as 59 per cent. titanium dioxide and 12 per cent. phosphoric pentoxide. Fluorine is present in quantities of 1 per cent. or more and sulphur to the same amount, but there is very little chlorine. The rutile as well as the ilmenite is recovered by concentration and is used mainly for the manufacture of arc-lamp electrodes.

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[1] F. L. Hess, Tin, tungsten, and tantalum deposits of South Dakota, *Bull.* 380, U. S. Geol. Survey, 1909, pp. 131–163.

[2] E. C. Andrews, The geology of the New England Plateau, *Records*, Geol. Survey, N. S. W., vol. 8, pt. 2, 1905, pp. 131–136.

L. A. Cotton, The tin deposits of New England, *Proc.*, Linnean Soc. N. S. W., vol. 34, pt. 4, November 24, 1909.

[3] T. L. Watson and S. Taber, The Virginia rutile deposits, *Bull.* 3-A, Virginia Geol. Survey, 1913.

T. L. Watson, Occurrence of rutile in Virginia, *Econ. Geol.*, vol. 2, 1907 pp. 493 504.

Wolframite.—As noted above, wolframite usually accompanies cassiterite in pegmatites, but only a small amount of the world's supply of tungsten is derived from these sources.

Columbite and Tantalite.[1]—These minerals are columbates and tantalates of iron and manganese ((Fe, Mn) $(Cb, Ta)_2O_6$). Their home is in the granitic pegmatites, from which the small quantities needed for incandescent lamps, electrodes and surgical instruments are derived. Large masses of columbite in black tabular crystals have been found in the pegmatites of the Black Hills, especially at the Etta mine. Mangano-tantalite, richer in tantalum is mined from similar sources in Western Australia. Columbite is not uncommon in many regions characterized by pegmatite dikes, such as Connecticut and Virginia. Strüverite ($FeO\ (TaCb)_2O_5.\ 6TiO_2$), isomorphous with rutile has been found in abundance at the Etta mine, South Dakota and, in places, in the Federated Malay States. The price of tantalum is about $18 per ounce.

Yttrium, Thorium, and Cerium Minerals.—Among the many rare earth minerals the following are the more important: thorite, (ThO_2); monazite (Ce, La, Y, Th) PO_4; gadolinite (beryllium-iron-yttrium silicate); allanite (cerium epidote); yttrialite (silicate of yttrium and thorium); euxenite (columbate and titanate of cerium metals and uranium); samarskite (columbate and tantalate of cerium metals and uranium). Some of these minerals, mainly monazite, are used extensively as a source of thorium salts in the manufacture of incandescent mantles; the yttrium minerals, like fergusonite and gadolinite, were used in the manufacture of Nernst lamps in which the incandescent part consisted of 25 per cent. yttria and 75 per cent. zirconia. The cerium minerals have a limited use for chemicals, etc., as well as for the manufacture of ferrocerium, an alloy emitting sparks when rubbed by a hard substance. The Welsbach incandescent mantles are coated by a substance containing 60 per cent. zirconia, 20 per cent. yttria, and 20 per cent. oxide of lanthanum.

At the present time incandescent mantles are said to contain 99 per cent. ThO_2 and 1 per cent. $CeCO_3$. The principal compound manufactured from the thorium minerals is thorium nitrate which formerly was worth about $2 per pound; the present

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[1] F. L. Hess, *Bull.*, 380, U. S. Geol. Survey, 1909, pp. 157–161; *Mineral Resources*, pt. 1, U. S. Geol. Survey, 1912, pp. 977–979.

price is about $8 per pound. Most of the thorium nitrate used in the United States was imported from Germany where it was manufactured from Brazilian monazite. In 1913, 119,000 pounds were imported. The small amount of cerium used is derived from monazite. Mesothorium, an element similar to radium and used for curative purposes and luminous paint, is also contained in thorium minerals and is extracted as a by-product.

All these minerals find their home in the granitic pegmatites and in the placers derived from them. To some extent they are also primary constituents of igneous rocks. In Scandinavia there are some celebrated occurrences, like those of Hitterö, in southern Norway, and of Ytterby, Korarfvet, Brodbo, and Finbo, in Sweden. One of the most renowned localities in the United States is Baringer Hill, 100 miles northwest of Austin, Texas; few other localities have yielded as large amounts of rare-earth minerals as this place.[1]

Baringer Hill is a low mound, about 100 feet wide and 250 feet long, preserved from erosion by its relative hardness. The country rock is a coarse, porphyritic granite of pre-Cambrian age, and the dike itself an unsymmetrical body or pipe. At the edge of the dike is pegmatite of the "graphic" variety 1 to 6 feet wide. The central part is made up of large individuals of quartz and feldspar, the latter being microcline and albite. In the center of the dike the quartz appears to be concentrated. Some of the feldspar crystals are several feet long. Vugs are lined with smoky quartz. The rarer minerals, some of which occur in large amounts, are fluorite, ilmenite, gadolinite, allanite, fergusonite, and polycrase—in short, a series of silicates, columbates, titanates, and uranates of cerium, yttrium, and other rare metals. There are also sulphides, particularly chalcopyrite, pyrite, sphalerite, and molybdenite, the last named being the most abundant. The rock contains no tourmaline, beryl, zircon,

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[1] W. E. Hidden and C. H. Warren, *Am. Jour. Sci.*, 4th ser., vol. 22, 1906, p. 515.

F. L. Hess, Minerals of the rare-earth metals at Baringer Hill, Llano County, Texas, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 286–294.

K. L. Kithil, Monazite, thorium and mesothorium, *Tech. Paper* 110, U. S. Bureau of Mines, 1915.

W. T. Schaller, Thorium minerals in 1916, *Min. Res.*, U. S. Geol. Survey.

W. T. Schaller, Zirconium and rare-earth minerals in 1916, *Idem.*

C. R. Böhm, Die Verwendung der seltenen Erden, Leipzig, 1913.

garnet, or cassiterite. The deposit is worked intermittently for the yttrium which its minerals contain. Some of the minerals show a marked radioactivity.

Monazite and Zircon.[1]—Both these minerals form accessories of granitic and monzonitic rocks; they also occur in pegmatites and apparently are formed in some veins developed at comparatively high temperatures. On the other hand, they are absent from veins formed nearer the surface or under conditions of lessened temperature and pressure.

Zircon ($ZrSiO_4$) occurs in considerable amounts in many placer deposits derived from the disintegration of granitic and pegmatitic rocks. In the miner's pan it is concentrated, with the gold, as a string of minute crystals of brilliant white, almost metallic luster. The best-known deposits are at Zirconia, near Green River, in Henderson County, North Carolina; from the decomposed croppings of a pegmatite dike at this locality many tons of zircon have been obtained. The value of the concentrated zircon sand is about 20 cents per pound.

A zirconium mineral,[2] at first thought to be baddeleyite (ZrO_2) has recently been found in Brazil in the Caldas District on the border of the States of Minas Geraes and Sao Paulo. It occurs in large quantities and is probably connected with pegmatite dikes in nephelite-syenite, and appears to consist of brazilite, a fibrous variety of ZrO_2, zircon and an unknown zirconium mineral with 75 per cent. ZrO_2, possibly a silicate but not identical with zircon. Much of this material is imported.

The principal use of zircon is as an excellent refractory material; it has also a very low coefficient of expansion. Zirconia is also used for incandescent mantles and zirconium as a steel hardening metal. Nickel-zirconium (cooperite) is another alloy used for high-speed tools. The value of zircon varies: $100 to $400 per ton have been paid.

Monazite, a honey-yellow to brown phosphate of cerium and cerium metals, with a varying percentage of thoria[3] and silica, is almost wholly recovered from placers where it often occurs with

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[1] D. B. Sterrett, *Mineral Resources*, U. S. Geol. Survey, 1908–1911.

J. H. Pratt, Zircon, monazite, etc., *Bull.* 25, North Carolina Geol. & Econ. Survey, 1916.

[2] W. T. Schaller, *Mineral Resources*, U. S. Geol. Survey, 1916, pp. 377–379.

[3] For commercial purposes the mineral should contain from 3 to 9 per cent. of thoria. See footnotes on p. 244. See also p. 770.

gold. The best-known occurrences are in South and North Carolina and in the Boise Basin of Idaho. The primary home of the mineral in these districts is in the granitic rocks and in pegmatized schist. The concentrates obtained in the sluices are cleaned in electro-magnetic separators. Large deposits of monazite, in part marine shore deposits of sand, are worked in Brazil and India, and the mineral is exported to Europe and the United States. The production in the United States varies considerably. In 1910 the output of concentrated monazite sand was 99,000 pounds, for which about 12 cents per pound was paid. From 1911 to 1915 there was no domestic production. In 1916, 1,200 tons of monazite sand was imported chiefly from Brazil.

Apatite.—Apatite associated with pyroxene (malacolite), hornblende, phogopite, titanite and much calcite occurs in many deposits in southern Ontario, near Kingston.[1] The deposits are said to be dikes of pyroxenite with segregations of apatite and calcite intruded in gneiss and limestone of pre-Cambrian (Grenville) age. It seems certain now that some of these pyroxene rocks are contact metamorphosed limestone (Fig. 254), but there are also dikes, though with what intrusions these dikes are associated is not certain. In part the apatite may be regarded as of pegmatitic origin. The apatite contains fluorine, instead of the usual chlorine. The mineral is well crystallized and of greenish color. In small quantities apatite also occurs here in veins in gneiss, and is then associated with pyroxene and scapolite. The output of Canadian apatite has ceased.

The Norwegian apatite veins[2] described in detail by Vogt now yield a small production. In some features they stand between the high temperature veins and the pegmatites. The minerals are chlorine, apatite, rutile, ilmenite, pyrrhotite, hornblende, enstatite, malacolite and specularite. The feldspar in the rocks adjacent to the veins are altered to scapolite indicating the presence of chlorine in the emanations.

Lithium Minerals.[3]—Among the alkaline metals lithium accompanies potassium in the pegmatites and appears in a series of

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[1] W. H. McNairn, *Trans.* Canad. Min. Inst., vol. 8, 1910, pp. 495–514.

F. D. Adams and A. E. Barlow, Geology of the Haliburton and Bancroft areas, *Mem.* 6, Geol. Survey Canada, 1910, p. 383.

[2] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1895, pp. 367–370, 444–459, 465–740.

[3] F. L. Hess, *Min. Res.*, U. S. Geol. Survey for 1909, pp. 649–653.

W. T. Schaller, *Min. Res.*, U. S. Geol. Survey for 1916, pp. 7–17.

minerals, the most common of which are lepidolite, or lithium mica (4 per cent. lithia); spodumene, or lithium-aluminum silicate allied to pyroxene (8 per cent. lithia); petalite, lithium-aluminum disilicate (5 per cent. lithia); triphylite, a lithium-iron-manganese phosphate (9 per cent. lithia); and amblygonite, a fluophosphate of aluminum and lithium (10 per cent. lithia). Spodumene and particularly amblygonite are the principal raw materials from which lithia salts are manufactured. These minerals have been mined in the Peerless and Etta pegmatite dikes, near Keystone, South Dakota. In 1916, 619 tons of this mineral were mined. Amblygonite also occurs in pegmatites near Pala, San Diego County, California. The use of lithium salts is small, chiefly for fireworks, medicine and certain storage batteries.

At the Etta mine attempts have also been made to mine the pegmatite for tin and columbite. The Etta deposit is a roughly circular mass of coarse pegmatite about 150 by 200 feet in extent. Spodumene crystals as much as 42 feet in length and having a cross section of 3 by 6 feet are found here. The list of primary minerals found at this remarkable locality is given below.[1] No topaz or axinite is present.

MINERALS FOUND AT THE ETTA MINE

Orthoclase
Albite
Microcline
Almandite
Grossularite
Andalusite
Muscovite
Biotite
Lepidolite
Petalite
Spodumene
Tourmaline
Epidote
Beryl
Titanite
Quartz
Zircon
Rutile
Spinel
Cassiterite
Corundum
Ilmenite
Columbite
Tantalite
Wolframite
Monazite
Amblygonite
Apatite
Triplite
Triphylite
Molybdenite
Arsenopyrite
Löllingite
Leucopyrite
Bismuth
Galena
Stannite

Cryolite.[2]—Cryolite ($3NaF.AlF_3$, with 12.8 per cent. aluminum) is a white to brown or even black mineral of which only one large deposit is known. The locality is Ivigtut, in west Greenland, close to the sea, where it occurs as a large mass having surface dimensions of 200 by 600 feet; it has been worked to a depth

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[1] F. L. Hess, *Bull.* 380, U. S. Geol. Survey, 1909, p. 149.

[2] N. V. Ussing, Denmark Geol. Undersög., 2d ser., No. 12, pp. 97–102.

R. Baldauf (and R. Beck), Ueber das Kryolith-Vorkommen in Grönland; *Zeitschr. prakt. Geol.*, vol. 18, 1910, pp. 432–446.

A. S. Halland, Cryolite and its industrial applications, *Jour. Industr. and Eng. Chemistry*, Feb., 1911, pp. 63–66.

of 150 feet. The cryolite occurs in a coarse granite and is undoubtedly to be classed as an unusual pegmatite mass. The coarsely crystalline mineral is associated with some crystallized siderite, galena, chalcopyrite, pyrite, fluorite, topaz, and ivigtite. The sulphides are said to contain a little gold. About 17,000 tons are produced annually, 4,000 tons being imported in the United States.

A pegmatite mass adjoining the cryolite contains the same minerals and also cassiterite in a coarse-grained aggregate of microcline, albite, and quartz. The cryolite is said to be intrusive into the granite and to effect many changes in it. The deposit is thus an unusually large magmatic concentration of fluorides.

Bauxite, the hydroxide of aluminum, is now used for the manufacture of the metal. Before the present methods of smelting aluminum were introduced the easily fusible cryolite was used for this purpose, and even now it is added to the charge to promote the melting. It is also used for enameling iron ware, and in making white Portland cement.

Precious Stones.[1]—The pegmatite dikes have always been famous as the source of gem minerals, which are valued for ornaments on account of their color, hardness, and brilliancy. Many of these beautiful crystals appear to belong to one of the later magmatic stages of consolidation and usually occur in druses of the rock. Among the most productive American regions are North Carolina, Maine, and San Diego County, California. The pink tourmaline of Pala and other places in San Diego County are famous and the crude output has an annual value of over $100,000. Accompanying this mineral are hiddenite and kunzite, the lilac colored gem varieties of spodumene.

Green tourmaline comes from Maine; emerald and aquamarine ($Be_3Al_2(SiO_3)_6$, both varieties of beryl, are found in pegmatites in North Carolina,[2] accompanied by quartz, albite, and tourmaline. Aquamarine of gem quality with much greenish beryl is found in pegmatite quartz of southern New Hampshire. The famous emeralds of Columbia,[3] at Muso, occur in carbo-

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[1] D. B. Sterrett, "Gems," *Mineral Resources*, U. S. Geol. Survey, 1908–1915.

[2] D. B. Sterrett, *Mineral Resources*, U. S. Geol. Survey, 1910, pt. 2, p. 865; 1911, pt. 2, pp. 1051–1052.

[3] J. E. Pogue. *Trans.*, Am. Inst. Min. Eng., vol. 55, 1917, pp. 910–934.

naceous, Cretaceous limestone, in calcite veins which seem to belong to the high temperature deposits rather than to the pegmatites.

Rubies are also found in pegmatite dikes. Most of the supply is derived from Burma, but some good stones have been obtained from gravels near pegmatite dikes in Cowee Valley, North Carolina.

Topaz is found in pegmatite as well as in lithophyse or rhyolite and in high temperature veins. Quartz of the clear, smoky and rutilated varieties is common in pegmatite.

Native Metals, Sulphides, and Arsenides.—The sulphides and allied minerals so abundant in fissure veins play a very subordinate part in the pegmatite dikes; nevertheless their occurrence is of great scientific interest, for the pegmatites form a transition between the magmas and many ore deposits connected with igneous rocks.

Gold in visible form is exceedingly rare; some instances are mentioned in Chapter I; additional and exact information is greatly desired. Spurr[1] states that gold occurs in pegmatite dikes or in quartz veins closely connected with them in the Yukon districts in Alaska and at Silver Peak, Nevada, but the quantities reported are, as a rule, small—at most 0.05 ounce per ton—and the assays are made without special precautions. I have no personal knowledge of any pegmatite nor of any quartz veins directly traceable into pegmatites that contain enough gold for profitable extraction. Spurr, however, describes important gold-bearing quartz veins at Silver Peak, Nevada, which are said to form transitions into alaskite or aplitic quartz-orthoclase rock.

In 1898 E. Hussak[2] described the Passagem lode, in Brazil, and regarded it as a gold-bearing pegmatite dike. Orville A. Derby[3] has lately reviewed the evidence and arrived at the following conclusions, which are quoted because they very likely apply to many similar occurrences:

> The Passagem lode presents evidence of three distinct processes of filling. . . . . . (1) An extensive fissure opened by stress. . . . . was closed by an invasion of pegmatite running off into clear quartz. At this stage the lode contained only the minerals characteristic of a

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[1] J. E. Spurr, Ore deposits of the Silver Peak quadrangle, Nevada, *Prof. Paper* 55, U. S. Geol. Survey, 1906, p. 49.

[2] E. Hussak, *Zeitschr. prakt. Geol.*, October, 1898, pp. 345–357.

[3] *Am. Jour. Sci.*, 4th ser., vol. 32, 1911, pp. 185–190.

granitic magma. . . . . . . (2) A subsequent stress . . . . . fractured this pegmatitic quartz, rendering it accessible to a pneumatolytic action . . . . . which filled its fissures with tourmaline and sericitized the feldspar of the pegmatite. (3) A third stress coming near the end of the second phase of the lode fractured the tourmaline filling and gave access to a pneumatolytic action characterized by sulphur, arsenic, metallic oxides, and metals (gold and silver), which filled the fissures of the lode, invading to some extent its pre-existing portions and probably also some of the adjacent and enclosed country rock.

Arsenopyrite, löllingite, galena, zinc blende, pyrite, and pyrrhotite have been reported from numerous localities in the granitic and syenitic pegmatites; there is not the slightest reason to doubt that they are here primary minerals, even if they belong to one of the later phases of magmatic consolidation. There are, however, no deposits known in which they are abundant enough to be mined.

Bismuth and bismuthinite are reported from many places, and they are said to be so abundant in certain pipes of pegmatite in the New England district of New South Wales as to have some economic value.

Molybdenite.[1]—Molybdenite is an accessory mineral in certain granites. It is common in many veins of the deep-seated class, more or less closely connected with pegmatites. It is also of frequent occurrence in contact-metamorphic deposits and in ordinary fissure veins, both in those formed at greater depth and in those deposited near the surface, but in the main is confined to deposits genetically allied to igneous rocks. In the pegmatite

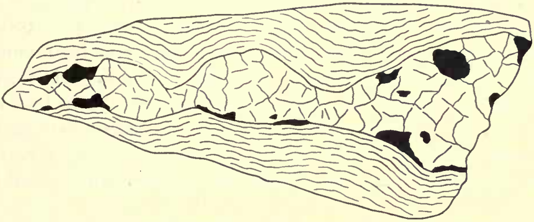

Fig. 261.—Molybdenite (black) along the borders of pegmatite dike in gneiss, Romaine, Quebec. One-half natural size. *After T. L. Walker.*

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[1] F. W. Horton, Molybdenum, its ores and their concentration, *Bull.* 111, U. S. Bureau of Mines, 1916.

and the abyssal veins the individual particles of molybdenite are often larger and sometimes well crystallized. In the deposits formed under conditions of less intense heat and pressure molybdenite usually appears as small or microscopic scales.

In Canada, in Quebec and Cape Breton provinces, some pegmatitic dikes contain enough molybdenite to be of economic importance (Fig. 261).[1] In Ontario, near Kingston,[2] molybdenite occurs with pyrrhotite in contact-metamorphic or pegmatitic deposits, similar to those which carry pyroxene, apatite, and brown mica (Fig. 254).

Pegmatites containing molybdenite occur in Washington and Hancock counties, Maine.[3] Some of them may have economic importance. One of the principal deposits occurs at Cooper, 22 miles southwest of Calais; here the molybdenite is especially associated with the more quartzose phases of the pegmatite. Fluorite in places accompanies the molybdenite.

Similar deposits are found in many parts of the United States but difficulties of concentration long confined the product to hand picked material. By aid of the oil flotation process the concentration of molybdenite has become possible and the mineral promises to become of great importance. So far most of the product has come from Ontario and British Columbia; in 1916, 84 tons of pure sulphide was produced from concentrates containing 60 to 70 per cent. of molybdenite. Pyrite, if present, must be eliminated. The concentrate is converted to ammonium molybdate, which is charged with iron in the electric furnace, yielding ferromolybdenum with 70 to 76 per cent. Mo, 0.1 per cent. S. and 3 per cent. carbon.[4] The price of concentrate with 90 per cent. MoS_2 was $0.75 to $1 per pound in January, 1919. Ores containing 1 per cent. MoS_2 can evidently be worked with profit.

At several places, for instance at Climax, Summit Co., Colorado, preparations are made to recover molybdenite, generally from

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[1] T. L. Walker, Report on the molybdenum ores of Canada, Canada Dept. of Mines, 1911, p. 64.

A. L. Parsons, Molybdenite deposits of Ontario, *Twenty-sixth Ann. Rept.*, Ontario Bur. Mines, 1917, pp. 275–313.

[2] E. Thomson, A pegmatitic origin for molybdenite ores, *Econ. Geol.*, vol. 13, 1918, pp. 302–313.

[3] Geo. O. Smith, *Bull.* 260, U. S. Geol. Survey, 1905, pp. 197–199.

F. L. Hess, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 231–238.

[4] H. H. Claudet, *Trans.*, Canad. Mining Inst., vol. 20, 1917, pp. 121–134.

aplite or pegmatite, sometimes from veins. New South Wales, Queensland and Norway also yield molybdenite.

Wulfenite ($CaMoO_4$), a yellow tabular mineral, is not uncommon in many oxidized deposits, the primary ore of which contained galena and molybdenite. Most of the small output of molybdenum in the United States has been derived from this mineral, though it is less desirable than molybdenite.

Aside from many chemical and industrial uses, the principal value of molybdenum lies in its steel-hardening qualities. Molybdenum steel, containing a very small amount of the metal, is used for rifle barrels, propeller shafts, and especially for high-speed steel-cutting tools.

CHAPTER XXIX

MINERAL DEPOSITS FORMED BY CONCENTRATION IN MOLTEN MAGMAS

CONSTITUTION OF MAGMAS AND THEIR DIFFERENTIATION AND CONSOLIDATION

General Features

Certain kinds of mineral deposits form integral parts of igneous rock masses and permit the inference that they have originated, in their present form, by processes of differentiation and cooling in molten magmas. The minerals are of simple composition and few in number; most prominent among them are magnetite, ilmenite, spinel minerals, cassiterite, pyrrhotite, chalcopyrite, molybdenite, löllingite ($FeAs_2$), arsenopyrite, corundum, platinum, and diamond. At some places the resulting deposits are large and rich, but as a whole they are of much less importance than those formed by aqueous solutions.

The characteristic feature of a deposit of this class is that it is a part of a body of igneous rock; the crystals of its minerals formed in the magma solution from which the rock crystallized, or in one similar to it. The associated gangue minerals are those which make up igneous rocks. Structures other than those of purely igneous origin should be absent. If there is evidence of metamorphism or metasomatic replacement, with the development of minerals like sericite, carbonates, chlorite, uralite, garnets,[1] or epidote, or bleaching or kaolinization, we must conclude that processes other than those of purely igneous origin have been active. Many igneous deposits have, at a later period, been subjected to influences producing alteration and their original characteristics may then have become veiled.

Some igneous deposits are simply parts of the rock, which contains disseminations of the useful mineral, like diamond in certain peridotites, and have then the form of that rock mass itself—a dike or a volcanic neck, for instance. In other deposits

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[1] Garnets are, however, also occasionally found in normal igneous rocks, but their primary or secondary nature is not difficult to establish.

the massive ore forms a dike, as in certain titaniferous magnetites. Or again the ore minerals may have become concentrated in parts of the igneous rock and form rudely tabular or wholly irregular, usually ill-defined masses in the rock. Unless the deposit is of large cross-section it can rarely be followed to great depth like a fissure vein, for the movements in a viscous magma facilitated the formation of irregular streaky or pasty masses—often termed "schlieren," after the German usage—rather than bodies persistent for long distances in a given direction.

These deposits represent extreme conditions in mineral formation; the temperature of basic surface lavas is considered to have been about 1,000° to 1,250° C., but the deep-seated granular rocks in which most of the igneous mineral bodies occur crystallize more slowly than lavas and in general at lower temperatures probably from about 575° to 1,000° C. The temperature before emission and consolidation may have been hundreds of degrees higher than the various figures given above.

Constitution of Magmas

In order to explain the genesis of the igneous deposits it is necessary to inquire into the nature of igneous magmas.[1] The magmas are not haphazard aggregates of elements and compounds. They are probably solutions of definite silicate compounds in one another (after the manner of a mixture of water and alcohol); certain oxides like silica, alumina, ferric oxide, and water may also be present, at least in a magma approaching the point of crystallization, and these silicates and oxides are freely miscible in any proportion. Diabases, leucite basalts, and similar rocks may be reproduced by dry fusion, but water is present in almost all magmas and is in fact necessary for the crystallization of a great number of rocks. That the magma is a solution is inferred from the lowering of the freezing-point as shown by the order of crystallization, and from the fact that some

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[1] J. H. L. Vogt, Bildung von Erzlagerstätten durch Differentiations Processe, *Zeitschr. prakt. Geol.*, 1893.

A. C. Lane, Wet and dry differentiation of igneous rocks, Tufts College Studies, vol. 3, No. 1, 1910.

Compare chapters on magmas and differentiation in "Natural history of igneous rocks," by Alfred Harker; "Igneous rocks," by J. P. Iddings; "Igneous rocks and their origin," by R. A. Daly; and "Geochemistry," by F. W. Clarke, all of which have been freely consulted in the preparation of this summary.

of the last residues of crystallization have the character of eutectic mixtures. Dissociation takes place to some extent and the magmas are electrolytes. Arrhenius and Königsberger believe that at high temperatures water must be a stronger acid than silica and that the latter exists as hydrates and basic silicates.

At the surface lavas emit water and other volatile substances and it is therefore concluded that before reaching the surface the magmas must be more or less heavily charged with such gases. When the magmas are forced to a higher level in the crust the pressure is diminished and a part of the volatile substances are liberated just as carbon dioxide escapes from soda water when the bottle is opened. Another part is still held, but most of that is doubtless expelled when crystallization takes place. The presence of water greatly affects the physical properties of the magma and especially increases the fluidity. Barus, for instance, obtained mixtures of various glasses with much water and these congealed at low temperatures as "solid solutions." Upon heating in the air, water is expelled and a pumice-like mass results which has a much higher point of fusion. Many pitch stones and obsidians, which contain much water, act in the same way.

Crystallization of Magmas

As in an aqueous solution, the successive crystallization of given minerals in these deposits is dependent upon their solubility in the rest of the magma and does not follow their temperature of fusion. When a salt dissolves in water the temperature of solidification is changed. Water freezes at 0° C., but an addition of sodium chloride to it depresses its melting or solidifying point many degrees. Alloys show the same behavior—for example, those with extraordinary low temperature of fusion, sometimes below 100° C. In the same way an igneous rock may become fluid at a temperature far below the average melting point of its constituent minerals, or even lower than the lowest of these.

On the other hand, no mineral can separate if the temperature, for a given pressure, is higher than the point of fusion of this mineral. Below this point crystallization takes place whenever the point of saturation of the solution for this mineral is exceeded. Some of its components will form isomorphous mixtures, but a part of it will remain in eutectic proportions, which differ according to the composition of the rock.

According to the empirical rule of Rosenbusch the separation of crystals in a silicate magma follows an order of decreasing basicity, so that at every stage the residual magma is more acidic than the aggregate of the crystals already separated out. This rule is subject to important exceptions, especially in basic magmas, but in the granitic and dioritic rocks the basic and difficultly fusible minerals, such as zircon, magnetite, apatite, ilmenite, and rutile, crystallize first. Then follow biotite, hornblende, and augite, or in general the magnesium and iron silicates, then the soda-lime feldspars, later orthoclase, and finally the residual quartz, which probably separates at about 800° C. The "mother liquor" of a granite thus becomes successively richer in silica. The "mineralizers," or the volatile substances, like boron, fluorine, and tin, follow the acidic rather than the basic constituents. The residue, in granitic rocks, is a solution rich in alkalies and silica, probably with water, which under certain circumstances may be a eutectic and may be pressed out of the partly consolidated magmas as if from a sponge and crystallize as pegmatites in fissures held open by the hydrostatic pressure of the fluids.

The order of crystallization of substances in a magma probably depends upon their relative abundance and upon their solubility in the eutectic.

Near the surface the order of crystallization is not entirely like that just outlined; there are usually two generations of crystals, and sometimes an older generation, of hornblende, for instance, may be resorbed and almost obliterated. In rock-forming minerals the volume of the crystallized substances is smaller than that of the corresponding fluid substance; their fusibility and also their solubility diminish with increasing pressure. A sudden release of pressure may then act as an increase of temperature and newly formed crystals may be remelted.

Much time has been given of late to the study of eutectic mixtures in rock magmas, especially by J. H. L. Vogt,[1] of Kristiania. In comparatively few magmas, however, does the residual part closely approach well-defined eutectic composition.

Melts of certain proportions of miscible salts will solidify together at a temperature lower than the point of congealing of each constituent. These are called eutectic mixtures, and their

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[1] J. H. L. Vogt, Die Silikatschmelzlösungen, Kristiania, 1903 and 1904.

minimum temperature, with its definite corresponding proportions, is called the eutectic point. The salts must be miscible; if not, they separate in layers. The salts must not act chemically upon one another, for if they do new compounds are formed. Finally, the salts must not be isomorphous, for then no eutectic point is possible; albite and anorthite, for example, crystallize together in all proportions and the melting points of the mixed crystals form a series with no eutectic depression.

The assumption of free miscibility is probably subject to some exceptions. Vogt, for instance, has brought evidence to show that sulphides are more soluble in basic than in acidic magmas and that the solubility increases at higher temperatures. This is, then, probably a case of limited miscibility, and Harker believes that the same may be true of alumina in the case of the association of corundum with peridotite magmas, and of the spinel minerals (like chromite) and the silicates.

Differentiation in Magmas

Differentiation, according to Iddings, means the separation of a homogeneous rock magma into chemically unlike portions. Modern views, based on field work and petrologic studies, include the belief that for each region, in each separate "magma basin," there is one essentially homogeneous magma from which by some process of differentiation the various rock types have been derived. In general it is thought that the primary magma was of intermediate composition and has been separated into basic and acidic forms, like basalts, latites, and rhyolites.

Lagorio,[1] in 1887, began the investigations on differentiation in his memoir on the nature of the glass-base or groundmass by calling attention to "Soret's principle," which states that when two parts of a solution are at different temperatures, the dissolved substance will be concentrated in the cooler portion.

This unequal cooling, it was thought, produced the heterogeneity in an originally homogeneous magma. The substances with which the magma was most nearly saturated tended to accumulate at the cooler points, leaving the warmer portions with an excess of the solvent material. There are many objections to this view. G. F. Becker showed that molecular diffusion would in a viscous magma require almost unlimited time. H. Bäckström has pointed out that although the action assumed

[1] Tsch. Min. u. pet. Mitt., vol. 8, 1887, p. 421.

by Soret's principle might cause changes in the absolute concentration, it would be powerless to alter the relative proportions of the dissolved substances. Absorption and assimilation of the substances contained in the surrounding rocks might alter the composition of the magma, and sometimes this undoubtedly takes place, although most intrusive contacts show little evidence of such assimilation. But such absorption would not, for instance, account for the occurrence of separated portions of titanic iron ore.

"Gravitative adjustment," advocated by J. Morozewicz and R. A. Daly, may play a considerable part in differentiation. According to this theory a great mass of magma, like a high column of salt solution, would separate into a denser substratum and a lighter upper part. The presence of mineralizing agents is also a factor of importance.[1] Certain constituents of the magma are more soluble in them than others and thus a magma rich in silica and alkali, containing many rarer metals, may have accumulated at the upper levels of a magma basin, while the basic portion of the magma remained below. G. F. Becker has indicated the possible importance of fractional crystallization, thus regarding the differentiation as a consequence of the general cooling process. Along the cooler walls the difficultly fusible minerals will separate first, and the process is aided by convection currents. The last portion of the fused mass to solidify will be the portion with lowest temperature of fusion and will therefore approximate a eutectic mixture. Along the walls of a dike basic minerals and iron ores may thus solidify, while the center will have a different composition. In the case of titanic iron ores the ilmenite probably crystallized first and settled to the bottom.

If the component parts of a slowly cooling magma are not miscible, a liquation will take place and the heavier parts, such as the molten sulphides, will settle to the bottom.

As pointed out by L. V. Pirsson,[1] the phenomena accompanying the eruption or intrusion of a magma are extremely complex, and no fully satisfactory explanation can be given of the process of differentiation. Liquation, influence of mineralizers, assimilation of wall-rocks, and pressure during consolidation are

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[1] C. H. Smyth, Jr., The chemical composition of the alkaline rocks, etc., *Am. Jour. Sci.*, 4th ser., vol. 36, 1913, pp. 33–46.

[2] *Bull.* 237, U. S. Geol. Survey, 1905, p. 196.

undoubtedly all of importance, but the most general cause of differentiation is probably fractional crystallization. The more closely the composition of a magma approaches eutectic ratios the less capable of fractionation it becomes. That crystals sink or float in melts and even in those of considerable acidity has been proved experimentally by Bowen.[1] He also concludes that crystallization controls differentiation of the sub-alkaline series of igneous rocks. Perhaps this is going too far for it seems probable that in deep magma basins there must have been some differentiation before the crystals began to separate out. The fact that magnetite and ilmenite sometimes form dikes suggests differentiation before consolidation.

PRINCIPAL TYPES OF DEPOSITS

Among the valuable minerals formed during the consolidation of magmas are diamond, platinum, chromite, ilmenite, magnetite, corundum, cassiterite, pyrrhotite, pentlandite, pyrite, chalcopyrite, molybdenite sperrylite, and apatite. A much more complex series of minerals is contained in the pegmatite dikes, which are described separately. For each kind of rock certain minerals are characteristic and most of the rocks are of the deep-seated type, crystallizing with granular structure.

Diamonds, chromite, platinum, and sometimes corundum are associated with peridotites, corundum also with certain nepheline syenites. Chalcopyrite, pyrite, pentlandite, and pyrrhotite follow the basic rocks, especially gabbros. Apatite and magnetite are connected with alkali-rich syenites; ilmenite and titaniferous magnetites with anorthosites (labradorite rocks) and gabbros; cassiterite with granite.

DIAMONDS[2]

Diamond is pure carbon, crystallizing in the isometric system. It is the hardest of all minerals and has a specific gravity of 3.52. Usually it is white or yellowish, but other pale colors are

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[1] N. L. Bowen, *Am. Jour. Sci.*, 4th ser., vol. 39, 1915, pp. 175–191; *Idem.*, vol. 40, 1915, pp. 161–185.

N. L. Bowen, The later stages of the evolution of igneous rocks, *Jour. Geology*, vol. 23, 1915, supplement, p. 91.

[2] Gardner F. Williams, The genesis of the diamond, *Trans.*, Am. Inst. Min. Eng., vol. 35, 1905, p. 440.

The diamond mines of South Africa, 2 vols. New York, 1905.

not uncommon. Rounded forms with aggregate structure are called bort, while the dark gray or black carbonado is granular and without visible cleavage. The last two varieties, mainly found in placers in Brazil, are used for drilling, set in steel. Diamond powder is used as an abrasive.

Until 1871, diamonds were obtained only from placers. They occurred thus in the Deccan mines in India, where the parent rocks may be pegmatite dikes, or perhaps serpentines. In Brazil, in the province of Minas Geraes, they occur in sands or gravels derived from conglomerates. Orville A. Derby says that they have no known relation to peridotites.

In the gold belt of California many small diamonds have been found—for instance, at Placerville and Cherokee—very probably derived from the extensive masses of serpentine occurring in the Sierra Nevada. Scattered diamonds have been found in the northern drift area in Indiana and Ohio. Along the Vaal River in South Africa fine stones are found which, according to some authors, cannot have been derived from the peridotites and serpentines. Some have thought that their original home was in the diabases or pegmatites of that region.

The only American occurrence of note is near Murfreesboro, Pike County, Arkansas,[1] where stones of good quality are found in decomposed peridotite. Up to 1911 a total of 1,260 stones, aggregating 574 carats, had been found. No later information is available.

In 1871 the Kimberley diamond field in South Africa was discovered, the first known occurrence of diamonds in primary rock. The district lies in the northern part of Cape Colony and the adjacent part of the Orange Colony. Another district centers at Jägerfontein, in the Orange Colony; still another at the Premier mine, near Pretoria in the Transvaal.

The diamonds in the Kimberley field are disseminated in

R. A. F. Penrose, The Premier diamond mine, Transvaal, S. A., *Econ. Geol.*, vol. 2, 1907, pp. 275–284.

D. B. Sterrett and W. Schaller, Gems and Precious Stones, *Mineral Resources*, U. S. Geol. Survey annual publication.

See also Stelzner and Bergeat, Die Erzlagerstätten, and F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol Survey, 1916, pp. 322–326.

[1] G. F. Kunz and H. S. Washington, Diamonds in Arkansas, *Trans.*, Am. Inst. Min. Eng., vol. 39, 1908, pp. 169–176.

D. B. Sterrett, "Gems," *Mineral Resources*, U. S. Geol. Survey, pt. 2, 1915, pp. 843–858.

volcanic necks, commonly called "pipes," of "kimberlite," a serpentine derived from peridotite and containing a little garnet and biotite. This rock breaks through the horizontal quartzitic sandstones, volcanic flows, and shales of the Karroo formation (Carboniferous to Triassic). The Kimberley pipe has been worked down to a depth of about 2,000 feet, its diameter being about 500 feet. Near the surface the serpentine was decomposed, forming the so-called "yellow ground," but at greater depth the not yet oxidized "blue ground" was met. The latter is now mined exclusively and the tough rock is allowed to slack on the surface for many months before it can be washed. The washing is effected on greased tables, the grease having the property of holding the diamonds while the other constituents of the rock are washed over it. The blue ground is a rock of dull, greasy appearance consisting chiefly of serpentine with abundant secondary carbonates and partly altered remnants of olivine enstatite, biotite, vaalite (a brown mica), garnet, diopside, chromite, ilmenite, diamonds, zircon, sapphire, kyanite, rutile, perofskite, apatite, and tourmaline. Apophyllite, chlorite, and calcite are secondary minerals. The blue ground is distinctly brecciated and many of the fragments of harder undecomposed rocks still remaining are rounded. The pipes are probably to be regarded as explosion vents and the rock filling them is undoubtedly of igneous origin. As to the diamonds, they are often crystallized as octahedrons, with convex or concave faces, but most of the crystals as recovered are broken. The color is white, yellowish, greenish, lilac, and even deep yellow. Small rounded masses with concentric growth, have also been noted and many dark-gray semi-transparent pieces are found.

The blue ground contains fragments of carbonaceous shale and Carvil Lewis thought that this rock might have furnished the material for the diamonds. According to later investigations this mode of origin is improbable. Stelzner, Bonney, and others have shown that the gem crystallized as an integral part of the magma. Stelzner mentions intergrowths of pyrope garnet and diamonds, and in the last few years Gardner F. Williams has collected specimens from Kimberley which show crystallized diamonds still partly enclosed by garnet and ferromagnesian minerals. Trübenbach and Bonney have also found that diamonds actually occur in fresh eclogite, a garnet-pyroxene rock closely allied to peridotite, from the Newlands mine, about 40 miles west of Kimberley. T. W. E. David recently described a ⅓-carat diamond from New South Wales, embedded in a solid matrix of hornblende diabase.

The production at Kimberley by the De Beers Company is from 1,500,000 to 3,000,000 carats per annum; the average price for the rough stone ranges from $7 to $12 per carat. The best ground is said to average about 1 metric carat (200 milligrams) per ton.

The deposit at the Premier mine is a pipe half a mile by a quarter of a mile in horizontal section penetrating the older Pretoria series of sediments. About 10,000,000 tons of material is treated annually, yielding about 2,000,000 carats. In this mine the largest diamond ever found was obtained; it is known as the Cullinan diamond and formed a broken octahedron 4 inches long and 2 inches wide; its weight was 3,024¾ carats, measuring 2 by 4 inches.

Much of the product from the South African mines is sold in the United States, the imports having a value of thirty to forty million dollars a year. Good cut stones sell at from $150 to $200 per carat.

Carbon is soluble in molten magmas and can crystallize from them. Small diamonds have been artificially produced in several ways, well summarized by F. W. Clarke. They have been obtained by dissolving carbon in molten iron, fused olivine, and other lime-magnesia magmas. The discovery of small diamonds in meteorites of iron or peridotitic rock is another fact clearly pointing to a magmatic origin of this mineral. Lately it has been shown by R. A. A. Johnson[1] that chromite, a mineral of the peridotite rocks, contains microscopic diamonds.

OTHER PRECIOUS STONES

Other precious stones contained in igneous rocks are sapphire (p. 807), garnet, and peridot.

Pyrope (magnesium-aluminum garnet) of the beautiful deep-red color which is necessary for gem quality is usually found in basic rocks of igneous origin. The garnets of Bohemia, obtained in washing a Cretaceous conglomerate, are probably derived from a serpentine. The diamond-bearing serpentine of South Africa contains pyrope of gem quality, called Cape

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[1] *Mem.* 22, Geol. Survey Canada, 1913, p. 83.

ruby. Almandite (iron-aluminum garnet) is not so extensively used. It occurs in granite and aplite, and also, as a product of metamorphic action, in crystalline schists. In the Navajo Reservation, Arizona, pyrope and peridot (yellowish-green olivine) are obtained as disintegration products of a basaltic rock.

PLATINUM AND PALLADIUM[1]

Nearly all the platinum of the world is derived from placers, mainly in the Ural Mountains in Russia, though smaller quantities come from Colombia, California, and New South Wales (p. 242).

Native platinum occurs as an alloy with others of the platinum group—osmium, iridium, palladium, ruthenium and rhodium. Native iridium, iridosmine, and other alloys are found with it.

An analysis of the crude platinum sand of California by Deville and Debray showed the following percentages: Platinum, 85.50; iridium, 1.05; palladium, 0.60; rhodium, 1.00; gold, 0.80; copper, 1.40; iron, 6.75; iridosmine, 1.10; sand, 2.95.

The platinum in the placers forms small rounded, also concretionary and knobby dark-gray pieces. Bright silvery scales of iridosmine occurs with it. In the Ural large pieces of platinum have been found, the largest weighing about 26 pounds.

Platinum has been found in primary deposits, but few of them are of economic importance. The modes of occurrence of platinum are as follows: 1. In placers; 2. Disseminated in peridotite and olivine gabbro, associated with chromite; 3. In magmatic deposits in basic rocks, associated with chalcopyrite and pyrrhotite (with palladium); 4. In small quantities in quartz veins; 5. In contact metamorphic deposits; 6. In traces in copper deposits of many kinds (with palladium); 7. Concentrated by

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[1] J. F. Kemp, *Bull.* 193, U. S. Geol. Survey, 1902.

C. W. Purington, *Trans.*, Am. Inst. Min. Eng., vol. 29, 1899, p. 3.

R. Beck, Lehre von den Erzlagerstätten, 1909, pp. 22–25.

D. T. Day, *Trans.*, Am. Inst. Min. Eng., vol. 30, 1900, p. 702.

F. W. Clarke, Geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, pp. 700–704.

C. W. Dickson, *Jour.*, Canadian Min. Inst., vol. 8, 1905, p. 192.

D. T. Day, W. Lindgren and B. M. Hill, Platinum, *Mineral Resources*, U. S. Geol. Survey, 1900–1917.

G. F. Kunz, Platinum, *Bull.* Pan American Union, Nov., 1917.

processes of oxidation in replacement ores of copper and gold in limestone (with palladium).

Daubrée showed, in 1875, that the Russian platinum is intergrown with olivine, pyroxene, and serpentine. Beck, on authority of S. Conradi, reports it in dunite rocks in Solowioff Mountain, in the Ural Mountains, and states that the metal forms zonar crystals of unquestionable magmatic origin lying between grains of chromite. Kemp has found platinum in dunite from the Tulameen River, British Columbia. The serpentines of this region also yield traces of platinum. The platinum of California is found only in placers, but the metal is believed to be derived from the serpentine areas so common in the Sierra Nevada.

In the nickel deposits at Sudbury, Ontario, which are considered of magmatic origin, platinum arsenide, sperrylite, probably accompanied by palladium arsenide, is formed as small silvery-white cubes intergrown with pyrrhotite and chalcopyrite. The same mineral was discovered at the Rambler mine,[1] Wyoming, in copper ores, mainly chalcopyrite and covellite, probably of igneous origin and forming a lens in a dioritic rock. Much palladium is also present.

Another rare mode of occurrence of platinum is in quartz veins, described by Bell,[2] from the southern island of New Zealand, from northern Finland, and from Canada.

A peculiar occurrence of platinum with wollastonite and grossularite in a contact-metamorphic rock has been reported from Sumatra.[3]

Palladium occasionally occurs alloyed with gold. E. Hussak[4] found in Brazil such palladium gold in a limestone close to the contact of an igneous rock. Platinum and palladium are also recovered in the electrolytic refining of copper bullion.

A gold-platinum-palladium deposit concentrated by processes of oxidation has recently been described by Adolph Knopf.[5]

The ore occurs in a lead-copper-gold replacement deposit in limestone in the Yellow Pine district, southern Nevada. The material, which contains the precious metals, is plumbojarosite, a sulphate of iron, lead and bismuth, and occurs in im-

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[1] S. F. Emmons, *Bull.* 213, U. S. Geol. Survey, 1903, pp. 94–97.

[2] J. M. Bell, *Econ. Geol.*, vol. 1, 1906, p. 749.

[3] L. Hundeshagen, *Trans.*, Inst. Min. and Met. vol. 13, 1903–4.

[4] *Sitz.-Ber.* Akad. Wiss. Wien. vol. 113, No. 1, July, 1904, pp. 1–88.

[5] *Bull.* 620, U. S. Geol. Survey, 1915, pp. 1–44.

portant amounts. It may average in ounces to the ton: Gold, 3.46; silver, 6.4; platinum, 0.70; palladium, 3.38. All metals are in minute division. The gold is rough, black and spongy, the palladium-platinum appears as microscopic black grains. The platinum metals were probably contained in the primary sulphides and most likely have been concentrated by solutions in which they were present in colloid suspension.

Production and Use.—At best the world's annual output of platinum, mainly from the Ural Mountains, is 400,000 troy ounces. War conditions have decreased the Ural production, while Colombia has attained 20,000 ounces. The output of platinum metals from all domestic sources is now (1917) about 7,400 ounces of which less than 1,000 ounces is from Pacific Coast placers; the rest comes from electrolytic copper refineries and oxidized ores; about one-half of this consists of palladium. The price of platinum per troy ounce has risen from $20 per ounce or less to the present fixed price of $105; palladium to $135, and iridium obtained in refining crude platinum or by picking out iridium grains from the concentrate to $175.

Platinum is used for jewelry, in dentistry, for chemical utensils, for spark devices and in the contact process for making sulphuric acid. Attempts are now made to save platinum by using alloys of tungsten, molybdenum, chrome-nickel and palladium.

Palladium, a silvery white ductile metal, but soluble in HNO_3, is used for silvering circles on surveying instruments, and other electroplating; also in dentistry and in alloy with gold as substitute for platinum.

Iridium, exceedingly hard and resistant, is used for hardening platinum; 5 to 20 per cent. are added. It is also employed for chemical and physical instruments, contact devices, etc.

Rhodium finds use with platinum for thermo couples. There is little demand for it.

Osmium is available in considerable quantities from the refining of platinum sand. Formerly it was used in incandescent lamps.

IRON AND NICKEL

Native iron occurs sparingly in some basalts. Large masses have been found in a basalt at Ovifak, west Greenland, where the natives used the metal for their tools and weapons. The iron contains from 2 to 3 per cent. of nickel and 3 per cent. of carbon and was long thought of meteoric origin. The basalt breaks through Tertiary beds containing coal, and it is believed by some that the metal was reduced from the rock by means of the coal. The nickel it contains militates against this view; more likely it was carried up from some deep-seated source by the basalt.

Awaruite ($FeNi_2$) is disseminated in gravels and also as small grains in the serpentine and peridotite of Red Mountain, on the south island of New Zealand, and is found also in sluice boxes for gold washing at Hoole Canyon, Yukon Territory.[1] A similar mineral, josephinite ($FeNi_5$), has been found in detritus in areas of serpentine in southwestern Oregon and a few other localities.

CHROMITE[2]

Chromite ($FeO.Cr_2O_3$),[3] a mineral of the spinel group and usually admixed with other spinel molecules, is an almost constant accessory of peridotites and of the serpentines derived from them and is often found in them, as accumulations large enough to be mined. The ore, more or less mixed with the rock, forms irregular bunches and masses along the contacts or in the interior of the intrusive masses; frequently also it forms ill-defined streaks or "schlieren." In part the ore may have a secondary origin, being developed together with magnetite during the process of serpentinization from primary chromite, picotite, chromium-diopside, etc. Late investigations, particularly those of Vogt, have shown that chromite in large masses

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[1] J. Keele, *Summ. Rept.*, Geol. Survey Canada, 1910, p. 257.

[2] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1894, pp. 384–393.

E. Glasser, Les richesses minerales de la Nouvelle Calédonie, *Ann. des Mines* (10), vol. 4, 1903, pp. 299–536.

J. H. Pratt, The occurrence, etc., of chromite, *Trans.*, Am. Inst. Min. Eng., vol. 29, 1899, pp. 17–39.

W. Glenn, The chromites of Maryland, *Trans.*, Am. Inst. Min. Eng., vol. 25, 1896, p. 481.

E. C. Harder, Some chromite deposits in western and central California, *Bull.* 430, U. S. Geol. Survey, 1910, pp. 167–183.

J. S. Diller, "Chromite," *Mineral Resources*, U. S. Geol. Survey, Annual publication.

[3] Theoretically, chromite should contain 68 per cent. Cr_2O_3 and 32 per cent. FeO, but Al_2O_3 and MgO are always present and the actual content of Cr_2O_3 is rarely more than 60 per cent. It is one of the most difficultly fusible of minerals, melting at 1,850° C. (A. Brun).

mainly represents purely magmatic separations in peridotite magmas. Vogt showed that the succession in the Norwegian deposits was chromite, olivine, and soda-lime feldspar, and in all cases the chromite appears to be the earliest consolidated constituent. Deposits in serpentine are often admixed with magnetite. It is noteworthy that during the weathering of peridotite few chromium silicates are formed, while nickel silicates often develop. A little chromiferous muscovite (mariposite), also ouvarovite or chrome garnet, as well as chloritic chromium minerals, in places accompany the chromite.

Copper minerals, especially chalcopyrite, are occasionally found with the chromite. The reddish niccolite (NiAs) has been found in serpentines and peridotites at Malaga, Spain.[1] The mineral is later than the chromite and according to R. Beck[2] cements crystals of augite.

Chromite ores should contain 45 to 50 per cent. Cr_2O_3. Deposits occur in many countries. An important occurrence has been found lately at Selukwe, Rhodesia, from which in 1910 44,000 tons of ore were produced. Until lately the largest supplies were received from Asia Minor, near Antiochia, Smyrna, and Brussa. These deposits are of unusual size. The Daghardy mine,[3] near Brussa, yielded annually 12,000 to 15,000 metric tons of ore. One of the masses from this mine was 70 meters long, 25 meters wide, and 20 meters deep. Much chromite is also exported from New Caledonia, where the ore occurs in part as residual masses which are concentrated, in part as "vein-like segregations" in serpentine. Smaller masses have been mined near Baltimore, Maryland, in North Carolina, in California, and in Oregon. Deposits of chromite in the serpentine areas of Quebec are now worked by concentration. Chromium is extensively used as a steel-hardening metal and also for the preparation of various salts, among which the bichromate of potassium is most important. Under war conditions prices for ore containing 50 per cent. Cr_2O_3 have risen to $80 per ton but even at that domestic sources have only contributed

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[1] F. Gilman, Notes on the ore deposits of the Malaga serpentines, *Trans.*, Inst. Min. and Met. (London), 1896, pp. 159–165.

[2] Erzlagerstätten, 1, 1909, p. 89.

[3] R. E. Weiss, *Zeitschr. prakt. Geol.*, 1901, pp. 249–262.

W. F. A. Thomae, Emery, chrome ore, etc., in Asia Minor, *Trans.*, Am. Inst. Min. Eng., vol. 28, 1899, pp. 208–225.

40,000 tons (1917) whereas in 1916, 116,000 long tons, besides much chromate, was imported.

ILMENITE OR TITANIC IRON ORE

General Features.—At many places in the world large masses of ilmenite ($FeTiO_3$, containing oxygen 31.6, titanium 31.6, iron 36.8), are found associated with more or less magnetite, olivine, pyroxene, and soda-lime feldspars. Petrographic research has long ago shown that ilmenite, with magnetite, is one of the earlier products of consolidation in magmas and is contained in almost all diabases, basalts, and gabbros; it occurs also in other less basic rocks, but the real home of ilmenite is in the rocks poor in silica. The larger masses of ilmenite are simply facies of the rock itself produced by concentration from the same magma. Near such masses the dark constituents first increase; finally the feldspar disappears and the ore-body consists of a mixture of ilmenite with ferromagnesian silicates. Vogt[1] first

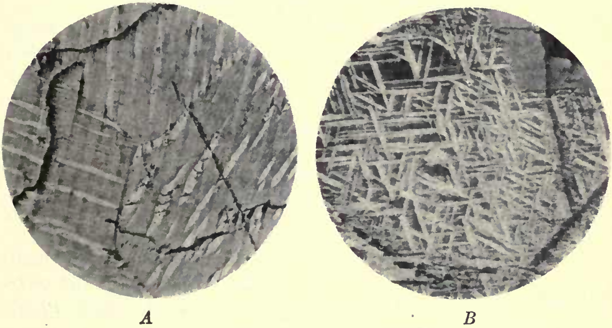

Fig. 262*A*.—Photomicrograph of polished section showing intergrowth of hematite (light); ilmenite (dark); St. Urbain, Quebec. Magnified 100 diameters.

Fig. 262*B*.—Intergrowth of magnetite (dark); ilmenite (light); a grain of olivine in upper right corner, Cumberland, Rhode Island. Magnified 180 diameters. *Both after C. H. Warren.*

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[1] Die Bildung der Erzlagerstätten durch Differentiation in basischen Eruptivmagmata, *Zeitschr. prakt. Geol.*, 1893, pp. 4–11, 125–143, 257–284; also in the same journal, 1894, pp. 381–399; 1900, pp. 233–242, 370–382; 1901, pp. 9–19, 180–186, 289–296, 327–340.

called attention to this well-defined group of ore deposits and explained its origin.

The ilmenite deposits, though large, have thus far been little utilized on account of difficulties in the metallurgical treatment; but these do not seem to be insuperable, and as it has recently been discovered that titanium confers valuable qualities of hardening on steel it may not be long before the ores will become important in metallurgy. During the last few years experiments in their utilization have been in progress in the United States.

Microstructure of Ilmenite.—The complex intergrowths of ilmenite with magnetite, rutile and hematite have been described lately by J. T. Singewald[1] and by C. H. Warren.[2]

Warren summarizes his results as follows: One type represented by occurrences at Miask, Arendal, and Snarum are homogeneous, though ilmenites from the latter two localities contain a great excess of Fe_2O_3 compelling the belief that there is a wide range of miscibility between the molecules $RTiO_3$ and Fe_2O_3 if these are really present. A second type presents an intergrowth of grains of homogeneous ilmenite and magnetites. The occurrences at Lake Sanford, Iron Mountain, Wyoming belong here. Many so-called titanic iron ores are magnetic.

A third type represented by St. Urbain (Fig. 262*A*) and Egersund, Norway, consists of a well-defined crystallographic intergrowth of ilmenite and hematite, possibly caused by the unmixing of an originally homogeneous solid solution.

A fourth type is illustrated by specimens from Cumberland, Rhode Island and Iron Mountain, Wyoming, showing an extraordinary minute regularly oriented intergrowth of magnetite and ilmenite (Fig. 262*B*). Warren believes that there exists a limited solid solution of the ilmenite and magnetite molecules, with a eutectic; and that ilmenite and hematite form a complete solid solution at higher temperatures, with an inversion interval and limited miscibility at lower temperatures.

Irregular Bodies.—The titanic iron ores form irregular masses or rather sharply outlined streaks in the central parts of gabbro

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[1] J. T. Singewald, The microstructure of titaniferous magnetite, *Econ. Geol.*, vol. 8, 1913, pp. 207–214; also *Bull.* 13, U. S. Bureau of Mines, 1913.

[2] C. H. Warren, *Am. Jour. Sci.*, 4th ser., vol. 25, 1908, pp. 12–38. *Am. Jour. Sci.*, 4th ser., vol. 33, 1912, pp. 263–277. The microstructure of certain titanic iron ores, *Econ. Geol.*, vol. 13, 1918, pp. 419–446.

or norite intrusives. The transitions to the country rock indicate that these masses have been formed by differentiation in the magma after the irruption in its present place. In these differentiated magmas ilmenite and magnetite have, as a rule, crystallized after the silicates. Where pyrite and spinel are present the order of crystallization is ferromagnesian silicates, pyrite, spinel, ilmenite (specularite), and magnetite. Probably little water was present and the temperature of consolidation was high, perhaps near 1,450° C., the fusion point of ilmenite, according to Brun. Vogt has shown that during the differentiation in a gabbro or norite magma a concentration of ferric oxide takes place, as well as of titanium, chromium, and vanadium; the lime, magnesia, and particularly silica diminish greatly, the silica to such an extent that the alumina and magnesia are forced to crystallize as corundum and spinel, both of which occur frequently in these deposits. Little sulphur or phosphorus is present.

Dikes.—Separated by a deeper-seated differentiation, veritable dikes of almost pure ilmenite may be injected into the prevailing rock, which then is usually an anorthosite.

Occurrences.—Vogt and Kolderup have described the Norwegian occurrences in norite and anorthosite in the great intrusive region in Ekersund; the largest body is 3 kilometers long and from 30 to 70 meters thick. Its composition is about 21 per cent. plagioclase, 41 per cent. hypersthene, and 38 per cent. ilmenite. At Routivare, in northern Sweden, a gigantic mass of titanic iron ore is included in a mass of highly altered gabbro, intruded in Cambro-Silurian strata. Some pyrrhotite is associated with the ore.

A large deposit at St. Urbain, in Quebec, is described by C. H. Warren.[1] Elongated, sometimes dike-like masses of ilmenite are inclosed in anorthosite. Specularite is intimately intergrown with ilmenite. Much rutile and blue grains of sapphirine ($Mg_5Al_{12}Si_2O_{27}$), also andesine, biotite, and spinel, are contained in some of the ore. Other varieties in which no rutile is present contain only 5 to 6 per cent. of other minerals.

J. F. Kemp[2] has described the large deposits in the Adirondacks of New York, near Elizabethtown and Lake Sanford.

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[1] *Am. Jour. Sci.*, 4th ser., vol. 33, 1912, pp. 263–277.

[2] Titaniferous iron ores of the Adirondacks, *Nineteenth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1898, pp. 383–422.

These ores are contained in a gabbro which is intrusive in a large "massif" of anorthosite. They are granular mixtures of magnetite and ilmenite with a maximum of 15 per cent. TiO_2, and form irregular or tabular masses presenting transitions to the country rock or appearing as distinct dikes. They contain plagioclase, pyroxene, olivine, hornblende, garnet, pyrite, apatite, spinel, and quartz, but are low in sulphur and phosphorus. At the principal locality several million tons of ore are probably exposed above the level of the lake.

Other deposits are known to occur in Minnesota[1] and Ontario.

In eastern Wyoming, at Iron Mountain,[2] a dike of almost solid ilmenite, in places 300 feet wide, breaks through anorthosite containing but little pyroxene and scarcely any ilmenite. This is a most remarkable instance of complete deep-seated differentiation of magmas. An analysis of the ore is as follows:

SiO_2	0.76	MnO	1.38
TiO_2	23.49	MgO	1.56
Al_2O_3	3.98	CaO	1.16
Cr_2O_3	2.45	P_2O_5	trace
Fe_2O_3	45.03	S	1.44
FeO	17.96	ZnO	0.47
			99.68
Total Fe			45.49

Some olivine, spinel, and magnetite are present as inclusions in the ore.

Concentrations of ilmenite with prevailing magnetite are not uncommon in gabbros and diabases, though rarely of economic importance.

Taberg, in southern Sweden, is of interest as one of the first places in which the existence of magmatic ore deposits was demonstrated, by A. Sjögren, Törnebohm, and Igelström. Taberg is a prominent hill 400 feet high, composed of norite. Toward the center the ilmenite and magnetite are greatly con-

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[1] F. J. Pope, *Trans.*, Am. Inst. Min. Eng., vol. 29, 1900, pp. 372–405.

N. H. and H. V. Winchell, The iron ores of Minnesota, Geol. Survey of Minn., 1891.

T. M. Broderick, Magnetic surveys of the magnetite deposits of the Duluth gabbro, *Econ. Geol.*, vol. 13, 1918, pp. 35–49.

[2] W. Lindgren, *Science*, new ser., vol. 16, 1902, pp. 984–985.

S. H. Ball, *Bull.* 315, U. S. Geol. Survey, 1907, pp. 206–212.

J. F. Kemp, *Zeitschr. prakt. Geol.*, 1905, p. 71.

centrated and form a mass of ore with some olivine, biotite and plagioclase. This body is 1 kilometer long and 450 meters wide and the material, although of low grade, has at times been smelted as iron ore; it contains about 6 per cent. titanic dioxide. A similar but smaller deposit contained in gabbro at Cumberland, Rhode Island, has been described by B. L. Johnson and C. H. Warren.[1] This ore contains 45 per cent. Fe_3O_4 and 10 per cent. TiO_2.

Influence of Pressure.—Kolderup has shown that, under dynamo-metamorphic influences, the ilmenite in deposits of this class changes to rutile, pyroxene is replaced by amphibole, garnet is developed, and new biotite and feldspar aggregates are formed.[2]

MAGNETITE

As iron is contained in some basic igneous rocks in great quantities and as even the average of this metal in all igneous rocks, according to F. W. Clarke, is 4.46 per cent., it might be expected that magmatic concentrations of magnetite would be abundant. This does not seem to be the case, however. In the early stages of consolidation of igneous rocks some magnetite is crystallized, together with other accessories, but the tendency to form silicates is strong and sufficient silica is usually available to take care of all the iron in the form of ferromagnesian silicates.

Commercially valuable deposits of magnetite as differentiation products of magmas have been found only in connection with syenites, syenite porphyries, and keratophyres, and here the magnetite is usually later than the ferromagnesian silicates and the feldspars. Certain magnetite deposits contained in ortho-gneiss at Sterling, New Jersey, are also believed to be caused originally by magmatic differentiation. Apatite frequently accompanies the magnetite, but sulphides are rare. Some basalts contain as much as 8 per cent. FeO and 4 per cent. Fe_2O_3; some gabbros and norites as much as 15 per cent. FeO and 1 or 2 per cent. Fe_2O_3. Magnetite requires 69 per cent. ferric oxide and 31 per cent. ferrous oxide. Any process of differentiation favoring the separation of magnetite thus requires a transfer of part of the iron to a ferric state.

Magnetite deposits are rarely found in surface lavas, doubtless

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[1] *Am. Jour. Sci.*, 4th ser., vol. 25, 1908, pp. 1–38.

[2] R. Beck, Erzlagerstätten, 3d ed., 1, 1909, p. 57.

because time before consolidation has not been long enough to permit of differentiation. Such concentrations in basalt have been described from Colorado[1] and a similar occurrence in andesite is on record from Nevada,[2] though in the latter case the author considers the magnetite and apatite to have been formed by replacement after consolidation.

The Iron Ores of Northern Sweden.[3]—The great magnetite deposits in the extreme northern part of Sweden, the largest in the world, have been studied lately by many geologists. For many years they remained unworked on account of their high percentage of phosphorus, which, since the invention of the Thomas process, is no longer objectionable.

The deposit at Kiruna forms a north-south ridge which is, as exposed at the surface, about 2.8 kilometers in length and rises 248 meters above Lake Luossajärvi, or 748 meters above the sea. The magnetite forms the highest part of the ridge and is from 32 to 152 meters wide. The total amount of ore proved above the level of the lake is said to be 265,000,000 metric tons, and the total, to a depth of 300 meters, proved by borings, about 740,000,000 tons. The present annual production from big open-cut workings is about 3,000,000 metric tons; the ore is exported to England, and to Germany.

The ore-body and surrounding formations are of pre-Cambrian age and dip steeply to the east (Fig. 263). In the footwall lies a syenite porphyry with fluidal structure and rich in soda (61 SiO_2, 8 iron oxides, and 6 to 7 Na_2O), the pyroxene of which is largely altered to amphibole, chlorite, and epidote. Magnetite is present in two generations, the later of which may surround the

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[1] H. S. Washington and E. S. Larsen, *Jour.*, Washington Acad. Sci., vol. 3, 1913, pp. 449–452.

[2] J. C. Jones, *Econ. Geol.*, vol. 8, 1913, pp. 247–263.

[3] Hj. Lundbohm, Kiirunavaara och Luossavaara järnmalmsfält, Sveriges Geol., Undersökn., Ser. C, No. 175, 1898. Refs. *Neues Jahrb.*, 1900, 1, pp. 79–80; *Zeitschr. prakt. Geol.*, 1898, pp. 423–426.

O. Stutzer, Die Eisenerzlagerstätten bei Kiruna, *Zeitschr. prakt. Geol.*, vol. 14, 1906, pp. 65–71, 137–142. Ref. *Econ. Geol.*, vol. 2, 1907, pp. 88–91.

P. Geijer, Igneous rocks and iron ores of Kiirunavaara, etc., Stockholm, 1910, pp. 278. Author's abstract, *Econ. Geol.*, vol. 5, 1910, pp. 699–718 (English).

P. Geijer, Studies on the geology of the iron ores of Lappland, Geol. För., *Förh.*, vol. 34, 1912, pp. 727–789 (English)

R. A. Daly, Origin of the iron ores at Kiruna, Stockholm, 1915, pp. 31.

feldspars and enter them along cleavage planes. The contact between the ore-body and footwall porphyry is apparently sharp, but shows in detail a narrow zone of transition due to small, sharply defined dikelets of magnetite in the porphyry. An ultimate product of this zone is a mixture of magnetite with green hornblende; it contains vugs filled with apatite, titanite, and magnetite. The deposit itself is free from vugs. The reddish quartz porphyry of the hanging wall is essentially similar in microscopic character, but contains about 71 silica, 5 iron oxides and 5 to 6 Na_2O, and has been classified as a quartz keratophyre. It contains fragments of magnetite ore and occasionally of the footwall porphyry. Above this hanging wall porphyry lie quartzites, clay slates, and conglomerates, with water-worn fragments of iron ores and keratophyre.

The ore is compact and fine-grained, consisting chiefly of magnetite and apatite in intimate intergrowth, apparently having crystallized together. In places it contains pyroxene. The ore is said to average 68 per cent. iron. The phosphorus is, as a rule, above 2 per cent., and some parts of the ore yield from 3 to 4 per cent. or even more of this substance. The sulphur is not above 0.05 per cent.; manganese not above 0.70 per cent.; a similar amount of magnesia is recorded, about 1.5 per cent. silica, 0.75 per cent. alumina, about 3 per cent. lime and 0.3 per cent. TiO_2. In places a fluidal structure of the magnetite and branching veinlets of apatite are observed in the ore.

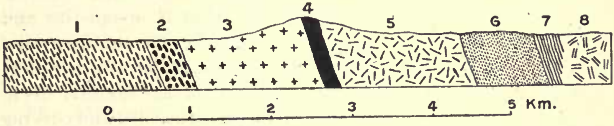

FIG. 263.—Schematic cross-section of iron deposit at Kiruna. 1, Soda greenstone; 2, Kurravara conglomerate; 3, syenite porphyry; 4, magnetite deposit; 5, quartz porphyry; 5 and 6, Hauki complex; 7, amphibolite; 8, quartz porphyry.

Early views on the genesis of the Kiruna deposit suggested pneumatolytic agencies. In 1898, Högbom proved its magmatic origin though he believed that the differentiation had proceeded in place. The investigations of Stutzer and the later monographic work by Geijer have shown plainly that the ore was differentiated from magmas in depth and that it has been brought to its present position in molten condition. The differences largely depend upon whether the porphyries are to be considered as effusive or intrusive. Stutzer held the latter and more probable view, regarding the syenite porphyry as the earlier rock followed by the intrusion of a dike of magnetite. On the east side the magnetite was later intruded by quartz porphyry which includes fragments of ore. Geijer, holding that the porphyries were extrusive, thought that the ore was erupted at the surface as a sheet of molten material while the flows were still horizontal. Lately R. A. Daly has again suggested a differentiation in place by gravitative settling of magnetite from the quartz-porphyry which he considers to be an intrusive sheet.

The great iron mines of Gellivare, a short distance south of Kiruna, produce about 1,500,000 metric tons of ore per annum.

The ore is principally mined in open workings and contains the same minerals as that of Kiruna—that is, magnetite and apatite—but it has a coarser grain. Locally it contains pyrite, chalcopyrite, fluorite, calcite, and zeolites. The ore is rudely schistose, conforming with the steep dip of the country rock, and forms large lenses, partly imbricating on parallel and curving strike lines.

The rocks are chiefly gneisses. The red gneiss is most common near the deposit and is traversed by many irregular veins of magnetite. It consists of albite with some quartz, chlorite, apatite, and biotite.

The reddish-gray gneiss is similar in composition but contains also soda-lime feldspar, microperthite, augite, and hornblende. Both rocks are rich in soda.

Dikes of acidic rocks, locally called granite but really quartz diorite or quartz keratophyre, cut across the ore-body.

The deposit at Gellivare has probably a similar origin to that at Kiruna. Epigenetic hypotheses are advanced by Lundbohm, von Post, and Löfstrand, the last two considering the deposit as a magmatic dike. This view is supported by the tectonic relationship; the ore-body is by no means confined to a single horizon in the gneiss. On the whole the analogy with Kiruna is very striking, though at Gellivare the rocks are clearly of intrusive origin. At both places the same genetic relations seem to exist; the earliest rock is rich in soda and of low to medium acidity, then follows an intrusion of magnetite-apatite rock, and lastly a quartzose soda-rich igneous rock was intruded.

Gellivare is, then, simply a dynamo-metamorphosed Kiruna.

The hematite deposits at Iron Mountain,[1] Missouri, are believed by Geijer[2] to be closely allied to the magnetite of Kiruna. He thinks the hematite may have been derived by alteration from magnetite and points to the association with apatite and the occurrence as dike-shaped masses in the pre-Cambrian porphyry.

The Magnetites of the Ural Mountains.—According to the recent investigations of Loewinson-Lessing and Jakowleff the magnetite deposits of Wyssokaia Gora and Goroblagodat, in the Urals, are products of differentiation in magmas, although at the former locality contact-metamorphic deposits also appear to be present. A summary of the Russian literature has been given by Beck.[3] In both places the igneous rocks are augite syenites; at Goroblagodat the ore forms streaks or "schlieren" in this rock; it has a brecciated structure, the magnetite cementing the augites and feldspar. The deposits show marked differences from the Swedish deposits just described in that they contain very little apatite and that the ores are not injected dikes, but perhaps rather products of differentiation in place.

The Magnetites of the Adirondacks.[4]—The eastern part of the Adirondack Mountains, in northern New York, contains a number of deposits of magnetite which have been worked since the early part of the last century and which still possess considerable economic importance. The total output is estimated at 40,000,000 long tons. The annual mine production in the last 25 years has varied from 1,000,000 to 2,000,000 long tons. The latter figure was reached in 1917. The more important operations are carried on in the Mineville district, but the deposits are spread over a large area. As some of the ores contain much apatite, magnetic concentration is used. The concentrates

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[1] C. W. Crane, The iron ores of Missouri, Missouri Bur. Geol. and Mines, 2d ser., vol. 10, 1912, pp. 107–144.

[2] *Econ. Geol.*, vol. 10, 1915, p. 324.

[3] Erzlagerstätten, 3d ed., 1, 1909, pp. 29–34.

[4] J. F. Kemp, Geology of the magnetites near Port Henry, *Trans.*, Am. Inst. Min. Eng., vol. 27, 1898, pp. 146–203.

David H. Newland and J. F. Kemp, Geology of the Adirondack magnetic iron ores, *Bull.* 119, N. Y. State Museum, 1908, p. 182.

D. H. Newland, Magnetites in the Adirondacks, *Econ. Geol.*, vol. 2, 1907, pp. 763–773.

contain 60 to 65 per cent. of iron, and a by-product of impure apatite is obtained which is used as a fertilizer. The tailings consist mainly of ferromagnesian minerals. The ores are extracted through shafts, the deepest of which, at Lyon Mountain, is 1,500 feet deep on the incline.

Until recently all the deposits in this region were considered as of sedimentary origin, for they are contained in crystalline gneissoid rocks, some of which are certainly metamorphosed sediments. In recent years, however, Kemp and Newland have shown that the ores stand in most intimate relationship to augite syenites.

The associated rocks include syenitic, granitic, and dioritic gneisses, garnetiferous schists, amphibolites, and crystalline limestones. The deposits considered of magmatic type occur in a belt of syenitic gneisses, in part also massive syenites and their pegmatites, whose igneous origin is well established. These rocks contain from 1.5 to 6.5 per cent. magnetite.

In the Archean sedimentary rocks, known as the Grenville series, are a number of smaller deposits, many of which contain pyrite as well as magnetite; in the rocks graphite, sillimanite, garnet, and scapolite have been noted. The genesis of these deposits is in doubt; they may be of sedimentary origin and subsequently metamorphosed.

The magnetites associated with undoubtedly igneous rocks appear as long lenses or pod-like bodies, often bent, curved, or folded, and have clearly participated in the general metamorphism of the country; at first they were probably tabular bodies. The ore lenses range in thickness from a few feet up to 25 feet or more, especially where curved or folded. In part the magnetite ore is pure, but more commonly it is mixed with the minerals that make up the wall-rocks, into which the ores often grade; these minerals are feldspar, quartz, pyroxene, and hornblende.

According to the percentage of phosphorus present the magnetites may be divided into low-phosphorus, Bessemer, and non-Bessemer grades. Apatite is usually present, and the non-Bessemer grade may contain as much as 10 per cent. of this mineral. While much of the ore yields 60 per cent. iron, there are large masses of ore with 50 per cent. iron or less that are suitable for concentration. According to Newland the lowest grade of workable milling ore carries about 35 per cent. iron. An average analysis of 65 carloads from pit 21 of the Mineville group of mines gave, in per cent., iron, 60.03; silica, 4.48; phosphorus, 1.635; sulphur, 0.021; and titanium, 0.12. The result of concentration from the "Old Bed" ore at Mineville in 1903 is shown in the subjoined table in percentages.

	Crude ore	Concentrates	First-grade apatite	Second-grade apatite
Iron	59.59	67.34	3.55	12.14
Phosphorus	1.74	0.675	12.71	8.06
Phosphorus as "bone phosphate."			63.55	40.30

The intimate association and intergrowth of the magnetite with the feldspar, augite, hypersthene, and hornblende of the augite syenite are considered by Kemp to prove its origin by magmatic differentiation. Syenitic pegmatites are also present and the processes of pegmatization are considered to have contributed to the formation of the ore; fluorite and titanite are often found in the ores.

CORUNDUM[1]

General Mode of Occurrence.—Corundum (Al_2O_3) has long been known as a product of regional and contact metamorphism; that it may also result from direct crystallization from a molten magma has been established by late investigations. Alumina is contained in all magmas, usually forming about 14 to 17 per cent. Certain syenites, nepheline syenites, and

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[1] The blue transparent corundum is called sapphire; the red transparent variety forms ruby; both varieties are valuable gems. Colorless, yellow, and green varieties also occur. The ordinary bluish or gray corundum is an inconspicuous mineral with good basal cleavage and great hardness, whence its principal use as an abrasive. Mixed with magnetite, mainly in metamorphic rocks, it is termed emery, the name being derived from Cape Emeri, on the island of Naxos.

J. H. Pratt, Corundum, *Bull.* 269, U. S. Geol. Survey, 1906.

A. E. Barlow, Corundum, *Mem.* 57, Geol. Survey Canada, 1915, p. 377.

G. P. Merrill, The non-metallic minerals, 1904, pp. 69–85; 2d ed., 1910, pp. 73–89.

J. Morozewicz, *Tsch. Min. pet. Mitt.*, vol. 18, 1898, pp. 22–83.

anorthosites may contain as much as 30 per cent. The peridotites, on the other hand, contain only from a fraction of 1 per cent. up to 10 per cent. of alumina. The corundum of magmatic origin is chiefly associated with those rocks of exceptionally high or low content of alumina, in which at the same time the silica is low.

By some observers the corundum of igneous rocks is regarded as due to recrystallization of included shale fragments. This view has been advanced by L. V. Pirsson in regard to the sapphire-bearing dike of Yogo Gulch, Montana. On the whole the differentiation theory fits the facts better.

Corundum is fusible only at 2,250° C. (Moissan), but it by no means follows that it crystallized from the magma at this temperature; Hautefeuille and Perrey showed, for instance, that alumina is soluble in a nepheline or leucite melt and that upon cooling the greater part of it crystallizes as corundum. Very likely it will be found that similar conditions would exist in melted olivine. Morozewicz has shown that when in a silicate melt the alumina is in excess of the ratio $RO:Al_2O_3::1:1$, corundum, spinel, sillimanite, or cordierite will form.

Corundum in Igneous Magnesian Rocks.—A long belt of magnesian rocks of probable pre-Cambrian age, mainly peridotites, gabbros, and norites, extends along the Appalachian region from Alabama to Massachusetts, and in these rocks corundum has been found in commercial quantities at a number of places.

In North Carolina and Georgia the mineral occurs as vein-like bodies from a few inches up to 15 feet in width at the contact of peridotite with gneisses and schists, in part also in the peridotite itself. Chlorite, enstatite, and spinel are associated with the corundum. Among the principal localities are the Laurel Creek mine, in Rabun County, Georgia; Corundum Hill, Macon County, and Webster, Jackson County, North Carolina. None of the southern localities have been productive in late years.

Deposits of emery, an impure corundum mixed with magnetite, are worked by open cuts near Peekskill, Westchester County, New York. The emery here occurs in the intrusive Cortlandt series of rocks described by G. H. Williams[1] and consisting of peridotites, norites, and diorites. The corundum and magnetite are, according to Williams, simply segregations in the norite, the constituent minerals of which occur even in the present

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[1] *Am. Jour. Sci.*, 3d ser., vol. 33, 1887, p. 135.

emery. Hercynite—an iron spinel—and garnet accompany the magnetite and the corundum.

Regional metamorphism easily changes many of these magnesian rocks to amphibolites and chloritic schists. During this process the corundum is apparently little affected. At Chester, Massachusetts, according to B. K. Emerson,[1] emery occurs for a long distance along a belt of amphibolite contained in sericite schists. The emery-bearing part of the schist is in places 12 feet wide and has been mined to a depth of 300 feet below the surface.

Corundum of gem quality is occasionally found in these deposits or in the gravels derived from them, but most of the sapphires obtained in the United States are derived from a different source. At Yogo Gulch, Fergus County, Montana, long dikes belonging to the monchiquite-camptonite series of lamprophyric rocks contain sharply crystallized rhombohedral sapphires of excellent quality. The decomposed rock is allowed to slack and is washed in sluice boxes. The deposits have been described by Weed,[2] Pirsson,[3] and Sterrett.[4]

Pale blue or greenish sapphires have also been mined on a commercial scale from the alluvial deposits extending for 20 miles along the Missouri River near Helena, Montana. According to G. F. Kunz the gems are derived from dikes of a vesicular mica-augite andesite, but the primary deposits have not been worked. The sapphires of Queensland are found in placers, associated with a basaltic rock. Those of India, Burma and Siam also occur in placers, and are derived from gneissoid or syenitic rocks.

Corundum in Syenite.—The most important deposits of corundum in Canada were discovered by W. F. Ferrier about 1896. The mineral occurs as an essential constituent of syenites and nepheline syenites and their pegmatites in Haliburton, Peterboro, Hastings, and Renfrew counties, Ontario.[5] The rocks form dikes or intrusive masses in gneisses and contain as much as 12 or 15 per cent. of bluish or gray, often roughly crystallized corundum, many of the crystals being 2 or 3 inches in diameter. The rock is quarried, crushed, and concentrated on tables.[6]

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[1] *Mon.* 29, U. S. Geol. Survey, 1908, pp. 117–147.

[2] *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1899, pp. 454–460.

[3] *Am. Jour. Sci.*, 4th ser., vol. 4, 1897, p. 421.

[4] *Mineral Resources*, U. S. Geol. Survey, 1907, pt. 2, p. 816.

[5] A. P. Coleman, *Eighth Rept.*, Ontario Bur. Mines, 1899, pp. 250–253.

[6] A. E. Barlow, *Op. cit.*

The production, formerly great, has now diminished. The value of corundum is $100 to $160 per ton.

Minor deposits of similar character occur in Gallatin County, Montana, in a syenite composed of orthoclase, biotite, and corundum.

Other Occurrences.—Corundum has also been found in anorthosites—rocks consisting principally of labradorite, or anorthite feldspar. According to T. H. Holland[1] corundum is abundant in India and representatives of the various classes already described are present.

In a number of occurrences of corundum in "gneiss" we have probably to deal with igneous rocks like syenite made schistose by pressure. Corundum is, however, unquestionably also developed by the contact metamorphism of limestone, probably by transfer of alumina from the magma. The largest known deposits of emery occur on the island of Naxos, in the Greek Archipelago, and near Smyrna in Asia Minor; they are contained in metamorphosed limestone. Corundum may also be developed by the regional metamorphism of clay shale and shaly quartzite. Many minor occurrences of this kind in the Southern States have been described by Pratt.

Production in the United States.—At present little or no corundum is produced in the United States. In 1916, 15,000 tons of emery at $8 per ton represented the domestic output. Imports of corundum and emery from Canada and other countries, in 1916, had a total value of about $241,000. In response to a great demand, artificial corundum (alundum) is now manufactured at Niagara Falls by fusing bauxite in the electric furnace.

Uses.—Corundum finds its principal use as an abrasive, wheels and abrading tools of all kinds being manufactured from the crushed material.

SULPHIDE ORES OF IGNEOUS ORIGIN

General Principles.—That sulphide minerals may crystallize from a magma has been ascertained beyond doubt, but the number of minerals which have this origin is limited to a few species, mainly pyrrhotite, pyrite, chalcopyrite, molybdenite, sphalerite, and pentlandite; arsenides like niccolite and sperrylite are also known. This view of the igneous origin of certain ores

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[1] A manual of the geology of India, Economic geology, part 1, Corundum, Calcutta, 1898.

has been gained comparatively lately, and largely by the labors of J. H. L. Vogt.[1]

While it is clear that sulphides are not freely miscible with silicate magmas,[2] Vogt has shown that the monosulphides are soluble to some extent in dry melts of basic character—that is, with much iron, calcium, and manganese; under favorable circumstances, particularly at temperatures of about 1,500° C., as much as 6 or 7 per cent. of these sulphides may become dissolved. But it is also to be noted that Vogt found but little solubility for the sulphides of copper, nickel, lead, and silver. Upon crystallization the sulphides always separate out first, as oldhamite (CaS), alabandite (MnS), troilite (FeS), and zinc blende (ZnS). These experimental results do not exactly correspond with those found in nature, for of the sulphides mentioned zinc blende is the only one encountered at all in igneous rocks, and the sulphides of slight solubility, like those of copper and nickel, are the most abundant. Later investigations by W. Wanjukoff[3] have shown that the sulphides of copper, nickel, iron, zinc and cadmium are soluble in basic slags to a notable degree and in the order of abundance indicated. Very likely the presence of mineralizers other than sulphur would increase this solubility as, in fact, already suggested by Vogt.

In the surface lavas which correspond most closely to dry melts primary sulphides are extremely rare, although grains of chalcopyrite are occasionally found. The sulphides of economic importance are almost wholly confined to the peridotites, norites, and gabbros, all rocks of deep-seated crystallization; the characteristic metallic association is iron, copper, nickel, platinum, and occasionally a little arsenic.

Many observers have stated[4] that the pyrrhotite and chalcopyrite which often occurs in basic rocks, more or less intergrown with magnetite, are of primary, magmatic origin. None has treated the subject better and more convincingly than E. Howe[5] who described the gabbro-norite and pyroxenite of the Cortlandt

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[1] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1893.

[2] J. H. L. Vogt, Die Silikatschmelzlösungen, pt. 1, Videnskabs-Selskabets. Skrifter, Math.-Naturv. Klasse, Kristiania, No. 8, 1903.

[3] *Metallurgie*, vol. 9, 1912, pp. 1–48.

[4] See W. Lindgren, Gold quartz veins of Nevada City and Grass Valley, *Seventeenth Ann. Rept.*, U. S. Geol. Survey, 1892, pp. 67–70.

[5] Sulphide bearing rocks from Litchfield, Ct., *Econ. Geol.*, vol. 10, 1915, pp. 330–347.

series. His conclusions are as follows: The extremely fresh rocks contain small amounts of pyrrhotite, pentlandite and chalcopyrite; there is no post magmatic alteration. The sulphides are as essentially magmatic as the silicates. Although most of the sulphides separated from solutions at an early stage of the cooling of the magma, small quantities continued to separate or to redissolve until the magma had nearly crystallized. The form and the interstitial relations of the sulphides seem to show that although they may have separated early from the silicates, they remained liquid until the silicates had crystallized. The sulphide bearing rocks are regarded as products of differentiation from magmas poorer in these substances.

The magmatic sulphide ores contain magnetite, pyrrhotite, pentlandite, chalcopyrite and bornite, but no gangue minerals other than the primary rock forming minerals. They contain no sericite, chlorite, garnet or epidote. The sulphides often corrode and embay the older silicates but without any indication of secondary substances. One often wonders what became of these dissolved portions. There is a certain succession among the sulphides. Where pyrite[1] is present it is often octahedral and etched depressions on its faces are sometimes filled with chalcopyrite. Pyrrhotite is the most abundant mineral. The usual succession of minerals as established by Tolman and Rogers,[2] is as follows: Silicates, magnetite, hematite, pyrrhotite, pentlandite, chalcopyrite and bornite. "Any change in this order, they state, is due to rearrangement subsequent to the magmatic period." Nevertheless it seems certain that, in these rocks, there is also magnetite, earlier than the silicates. It will be remembered that Vogt in his studies on silicate melts frequently obtained two generations of this mineral.

At times it becomes extremely difficult to hold closely to the above mentioned definitions and rules. Most deposits of magmatic sulphides are not uniform disseminations but rather marginal concentrations, quite plainly injected after the surrounding rock had crystallized, or else they are accompanied by gangue materials like chlorite, quartz, garnet, epidote, etc. Such deposits were probably formed by highly concentrated sulphide

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[1] Pyrite as a magmatic mineral is not accepted by all authors. See Tolman and Rogers, A study of the magmatic sulfid ores, Stanford Univ. Pub. Univ. Ser., 1916, p. 69.

[2] *Idem.*

melts with water and other mineralizers. When the dominating influence is no longer the straight melt but is instead the concentrated and controlling gases we may perhaps employ the much abused term pneumatolytic action, which again grades into hydrothermal action.

For the plainly magmatic ores, be they sulphide or oxides, Graton and McLaughlin[1] have suggested the term *orthotectic* while for the slightly later processes in which the strictly magmatic influences were modified by mineralizers they would use the expression *pneumotectic.* Naturally, the boundaries are sometimes indistinct. The Sudbury ores would be considered pneumotectic.

In the orthotectic deposits the temperature may have been very high but in the subsequent though still magmatic phases temperatures as low as 400° to 500° C. may have been reached. We are not well informed on this subject.

The magmatic sulphide ores have lately been discussed by W. H. Goodchild from a physico-chemical standpoint.[2]

Types of Deposits.—Some of the magmatic sulphide deposits are simply basic rocks abnormal in containing much pyrrhotite, chalcopyrite, and pentlandite. Other occurrences are clearly related to contacts and bear evidence of later magmatic injection. There is still another class in which the magmatic origin is only dimly perceived on account of the metamorphic changes which the rocks have undergone. The basic igneous rocks are easily modified by pressure and more or less schistose amphibolites are developed, which besides amphibole contain garnet, quartz, epidote, and chlorite. Any primary sulphide segregations contained in such rocks will be correspondingly affected and a new type of deposit of metamorphic appearance will result; the sulphides themselves apparently undergo little change.

Sulphides in Peridotites and Gabbros.—E. S. Bastin[3] has described a rock from Knox County, Maine (Fig. 264), which shows convincingly the magmatic origin of sulphide ores. This rock, which occupies a small area surrounded by drift, consists of 60 per cent. olivine, 21.53 per cent. pyrrhotite, some andesine-labradorite feldspar, hornblende, and magnetite, 1.03 per cent. chalcopyrite, and pyrite, biotite, and spinel. The analysis

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[1] *Econ. Geol.*, vol. 13, 1918, p. 85.

[2] *Mining Mag.* (London), Jan.–June, 1918.

[3] *Jour. Geology,* vol. 16, 1908, pp. 124–138.

shows 0.94 per cent. nickel oxide, and the material is thus practically a very low grade ore. The constituents are intergrown, showing simultaneous crystallization except that the magnetite, enclosed in olivine, is the earliest mineral separated; the chalcopyrite is associated irregularly with the pyrrhotite. There has been some serpentinization but not enough to obscure the relations. The complete analysis is as follows:

FIG. 264.—Thin section of olivine corroded by pyrrhotite and chalcopyrite East Union, Maine. Magnified 15 diameters. *After E. S. Bastin.*

SiO_2	28.04	MnO	0.24
Al_2C_3	3.51	Fe_7S_8	21.53
FeO	14.95	NiS	0.94
Fe_2O_3		CoS	0.03
MgO	21.97	$CuFeS_2$	1.03
CaO	1.78	H_2O+	2.54
TiO_2	0.20	H_2O-	1.48
P_2O_5	0.04	CO_2	1.01
K_2O	0.08		
Na_2O	0.28		99.65

In the deposit at Mittel-Sohland, in Saxony, described by Beck,[1] the sulphides form a rather rich nickel ore. They occur in an olivine diabase of gabbroic habit, containing, in order of deposition, magnetite, ilmenite, silicates, pyrrhotite, and chalcopyrite. The ores lie along the contact between the diabase and a granite

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[1] *Zeitschr.* Deutsch. geol. Gesell., 1903, pp. 296–331.

and extend in a belt about 2 meters wide for a distance of 150 meters, gradually merging into normal diabase; the granite contains disseminated sulphides close to the contact. Beck believes that the ores were introduced after the consolidation of the rock.

Vogt has described the numerous Norwegian occurrences in great detail. The ore-bearing intrusives are norites or allied rocks, often with biotite and brown hornblende, and are intruded in pre-Cambrian gneiss. In part the gabbros are pressed to amphibolites. The nickeliferous pyrrhotites occur largely along the contacts. They contain little copper and only 1 to 1.5 per cent. nickel. In the amphibolitized deposits considerable migration has taken place. Garnet is formed along the streaks of pyrrhotite. The hornblende is in part transformed to bluish-green amphibole.

A similar occurrence is that of Lancaster Gap, Pennsylvania, described by Kemp, where the nickeliferous pyrrhotite lies along the contacts of a mass of amphibolite, contained in mica schist. Much nickel ore was mined here up to 1893.

Many copper deposits in amphibolite are really dynamometamorphosed forms of such magmatic deposits as have been described above. The examination of several such small deposits in Colorado[1] led to this conclusion. Deposits at Sedalia and Turret, in Chaffee County, are basic dikes in a pre-Cambrian sedimentary series contact-metamorphosed by later granitic intrusion and still later altered to amphibolite. Chalcopyrite, zinc blende, and magnetite, are intergrown with bluish-green amphibole, garnet, spinel, and labradorite. The diabase is probably a differentiated offshoot from a neighboring large mass of coarse diabase.

Sudbury, Ontario.[2]—The nickel deposits of Sudbury now yield

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[1] W. Lindgren, Notes on copper deposits in Colorado, *Bull.* 340, U. S. Geol. Survey, 1908, pp. 157–174.

[2] The literature is extensive; only a few articles are cited here.

C. W. Dickson, *Trans.*, Am. Inst. Min. Eng., vol. 34, 1904, pp. 3–67.

A. E. Barlow, Report on the nickel and copper deposits of the Sudbury mining district, *Fourteenth Ann. Rept.*, Canada Geol. Survey, pt. H, 1904.

W. Campbell and C. W. Knight, *Econ. Geol.*, vol. 2, 1907, pp. 350–366.

A. P. Coleman, The nickel industry, Canada Dept. Mines, 1913.

E. Howe, *Econ. Geol.*, vol. 9, 1914, pp. 505–522.

C. F. Tolman and A. F. Rogers, *Op. cit.*, pp. 21–37.

A. P. Coleman, *Econ. Geol.*, vol. 12, 1917, p. 427.

the larger part of the world's production, and the once pre-eminent ores of New Caledonia are now of decreasing importance. Minor amounts of nickel are derived from deposits in Norway and elsewhere. No important nickel deposits have been found in the United States. The production from Sudbury in 1910 was 18,600 long tons of nickel while in 1917 the output exceeded 40,000 tons and was derived from 12 mines, which yielded about 1,500,000 tons of ore. Besides nickel the ores contain an important percentage of copper, with a little gold, silver, palladium and platinum. It is not necessary to specify the uses of nickel; they depend on its properties of toughening, whitening, hardening, increasing the elasticity and preventing the oxidation of certain alloys. Nickel steels are the most important of all alloy steels.

The geology of the region is complicated. On a basement of Keewatin greenstone and the Sudbury quartzite (lower Huronian) rests a syncline of upper Huronian or Animikie rocks including conglomerate, tuff and slates. This syncline is 36 miles long and 16 miles wide (Fig. 265). Between the basement and the Animikie there is intruded a thick sheet of igneous massive rocks which may be of Keweenawan age and is also referred to as the "nickel eruptive" on account of its unquestionable connection with the nickel deposits. This sheet is strongly differentiated, supposedly by gravitative settling of crystals in the magma. It grades from a norite or hypersthene gabbro in the lower parts to a micropegmatitic granite in the upper parts. Even in the lower section there is a certain amount of "acid extract" of micropegmatite between the other constituents.

Granitic rocks are intruded in the basement and there are even some granite dikes in the norite at the Murray mine and possibly at the Creighton mine.

The deposits are found (1) as rudely tabular masses at the contact of the norite with the basement rocks (marginal deposits); they dip 30° to 60° toward the center of the syncline (Creighton, Gertrude, Murray mines). (2) As mineralized dikes or "offset deposits" in the basement rocks. These are

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A. P. Coleman, The Sudbury laccolitic sheet, *Jour. Geology*, vol. 15, 1917, p. 252.

W. G. Miller and C. W. Knight, Nickel deposits of the world, Royal Ontario Nickel Comm., Toronto, 1917.

A. M. Bateman, *Econ. Geol.*, vol. 12, 1917, pp. 391–426.

M. A. Dresser, *Econ. Geol.*, vol. 12, 1917, pp. 563–580.

steep, irregular or columnar and have been followed to depths of 1,200 or 1,400 feet (Copper Cliff and Victoria mines).

The ore minerals are pyrrhotite, pentlandite and chalcopyrite with occasional magnetite, pyrite, sphalerite, sperrylite ($PtAs_2$), polydymite (Ni_4S_5). The order of succession as established by Tolman and Rogers is silicates, magnetite, pyrrhotite, pentlandite and chalcopyrite. Pentlandite often forms veinlets in pyrrhotite and can easily be distinguished from the latter in polished section by etching with hydrochloric acid. The precious metals seem to follow the chalcopyrite and occur most abundantly in some of the "offset deposits" like Victoria and Vermilion. The ores of the Creighton mine are the richest, containing in per cent. about 4 nickel and 1.4 copper. Other deposits yield poorer ore with 2 nickel and 0.8 copper. The values of the precious metals aggregates $1 to $2 per ton.

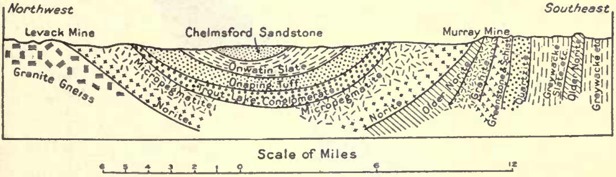

FIG. 265.—Section across Sudbury syncline showing intrusive norite sheet. *After A. P. Coleman.*

The best developed marginal deposit is at the Creighton mine where it has been followed to a depth of 2,000 feet on the incline (Fig. 266). It lies between granite and norite but as in other marginal deposits the ore is a breccia or mass of subangular boulders of almost barren norite, cemented by the ore minerals, which often form a hard crust on the rock. The sulphides also enter norite and granite as veinlets. The massive ore contains abundant remnants of partly replaced rock minerals and in the poorer ores the sulphides have corroded and replaced the rock minerals. However, the interpretation of the facts observed under the microscope varies considerably. No gangue minerals are formed. The ore shoots are from a few feet to 150 feet thick. Note similarity of Fig. 266 to sections of Norwegian deposits given by Vogt

In the "offset deposits" the relations are similar although we here find various gangue minerals such as chlorite (Copper Cliff), epidote and quartz indicating somewhat different solutions and probably lower temperatures.

The earlier view of a gravitative settling of the sulphides in the norite sheet has given way to the theory of an injection of sulphide magma more or less charged with mineralizers along certain brecciated or fractured zones. Graton's term of "pneumotectic deposits" (p. 811) is applicable to Sudbury and in places the deposits even show transitions to high temperature veins. In some respects they present strong similarity to the "injected pyritic deposits" (p. 818). In minor part they may have been formed by direct magmatic segregation from the nickel eruptive, but in greater part they have been formed at the end of the magmatic period by replacement of the silicates by a very liquid melt charged with sulphides and developed by differentiation in a magma reservoir in depth. With these views agree, wholly or partly, Tolman and Rogers, Howe, Bateman, and even Coleman and Knight. That the nickel ores are genetically connected with the norite admits of no doubt.

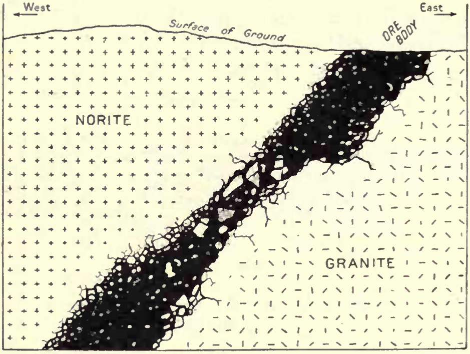

FIG. 266.—Generalized section through Creighton ore-body, Sudbury, Ontario, extending to eighteenth level. The norite contains blebs of ore about size of peas, for 2000 feet beyond the ore-body. *After C. W. Knight.*

Cape Colony.—A deposit at Insizwa, Cape Colony, similar to that of Sudbury, has been described by A. L. Dutoit and W. H. Goodchild.[1] Pyrrhotite, chalcopyrite, and pentlandite, with a little bornite and niccolite, occur in a thick sheet of gabbro or norite, at its contact with underlying sediments. The ores contain also a little platinum, osmiridium, gold, and silver. The sulphides separated from the magma at the end of the period of consolidation and corrode idiomorphic olivine (Fig. 267), pyroxene, and feldspar.

Bornite Deposits.—Bornite is occasionally recorded as a minor constituent of pegmatite dikes and sometimes occurs in the deep vein zone. A small but remarkable bornite deposit in an igneous rock, described by E. Ritter[2] and lately by E. S. Bastin and J. M. Hill,[3] occurs at the Evergreen mine, Gilpin County, Colorado. Dikes of a monzonitic rock, in part brecciated, contain, intergrown with the primary minerals, bornite and chalcopyrite, also garnet, calcite, and wollastonite. All these minerals are contemporaneous with the ordinary rock minerals. This seems to be a case of digestion of material from calcareous rocks; possibly the sulphides are also of foreign origin, perhaps derived by absorption from an older deposit; an origin by direct differentiation is, however, not unlikely. The ore extracted contained 3 per cent. copper and $5 in gold and silver per ton.

The remarkable and rich bornite deposits of Ookiep in Namaqualand, Cape Colony, have been regarded as magmatic by Stutzer, Kuntz and other authors. However, here too, the idea of fractured zones directing or enriching the ores has been brought forward.[4]

Tolman and Rogers[5] have lately examined these ores and conclude that they are of typical magmatic origin. Magnetite, hematite, chalcopyrite and bornite replace the silicates in hypersthenite and diorite.

The Engels Mine, Plumas County, California, contains disseminated bornite and chalcocite in a mass of norite. Some

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[1] *Fifteenth Ann. Rept.*, Geol. Comm., Cape of Good Hope (1910), 1911, pp. 110–142.

W. H. Goodchild, The economic geology of the Insizwa range, *Trans.* Inst. Min. and Met. (London), 1916.

[2] *Trans.*, Am. Inst. Min. Eng., vol. 38, 1907, pp. 751–765.

[3] *Econ. Geol.*, vol. 6, 1911, pp. 465–472.

[4] A. Schenk, Vortr. *Zeitschr.* d. d. Geol. Ges., 53, vol. 4 H, 1902, p. 64.

[5] *Op. cit.*

authors[1] have attributed the bornite to strictly magmatic processes while the chalcocite was believed deposited by concentrating thermal waters. Later, Tolman and Rogers[2] regarded the bornite as deposited by the aid of mineralizers at a later magmatic stage.

Still later Graton and McLaughlin[3] classified the deposit as of pneumatolytic origin with amphibole, albite, tourmaline, magnetite as gangue minerals and bornite and chalcopyrite as ore minerals. This was followed by hydrothermal action producing chlorite, sericite, epidote and bornite, and finally, zeolites. The chalcocite is regarded as a product of descending meteoric waters.

INJECTED PYRITIC DEPOSITS

General Features.—Some deposits in which the ore consists mainly of solid pyritic minerals present features which can hardly be explained otherwise than by actual injection of molten sulphides, perhaps to be considered as residual solutions from adjoining intrusive bodies. A. Bergeat[4] first summarized these peculiar occurrences, among which, it must be confessed, are

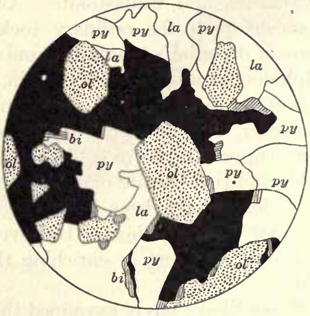

FIG. 267.—Thin section of olivine norite, Insizwa mine, Cape Colony. Magnified 18 diameters. Black areas, pyrrhotite; *ol*, olivine; *py*, pyroxene; *bi*, biotite; *la*, labradorite. *After A. L. Du Toit.*

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[1] H. W. Turner and A. F. Rogers, *Econ. Geol.*, vol. 9, 1914, pp. 359–391.

[2] A study of the magmatic sulfid ores, Stanford Univ. Pub., Univ. Ser., 1916.

[3] L. C. Graton and D. H. McLaughlin, *Econ. Geol.*, vol. 12, 1917, pp. 1–38.

[4] A. W. Stelzner and A. Bergeat, Die Erzlagerstätten, vol. 2, pp. 964–987.

some of the most enigmatic of ore deposits. The examinations of the deposit at Bodenmais, in Bavaria, by J. Lehmann and E. Weinschenk appear to have led up to the definite view suggested above, and since then a number of other occurrences have been added to this class, which presents strong points of similarity to the Sudbury deposits as interpreted above.

These ores are usually enclosed in schist or gneiss and that they originated by metasomatic replacement of limestone appears to be out of the question, though it must not be overlooked that the same was thought once of the ores at Ducktown, Tennessee, which now have been shown by W. H. Emmons to be replacements of small limestone lenses in a prevailing schist formation (p. 751).

That fluid sulphides may penetrate silicate rocks in veinlets and corrode the various primary minerals, like augite, has been shown in interesting experiments by O. Stutzer,[1] and by previous observations by von Cotta on the brickwork of old lead furnaces. In Stutzer's experiments the sulphide veinlets of pyrrhotite, zinc blende, and galena penetrated the rocks along minute cracks and along the cleavage planes of the minerals. In gabbros the pyroxene grains were corroded, in a manner similar to that noted in the ores of Sudbury. The sulphide melt would probably be under high pressure and would force its way into the adjoining rocks. Deposits of this kind are decidedly rare. The igneous rocks near whose contact injected deposits lie are of many kinds, not always of basic character.

The minerals of the ores include magnetite, pyrrhotite, pyrite, zinc blende, chalcopyrite, and rarely galena. The associated gangue minerals surely indicate high temperature and are present in scant quantity; they are quartz, orthoclase, plagioclase, amphibole, hypersthene, biotite, cordierite, spinel, especially zinc spinel, and garnet. The various minerals are practically contemporaneous. The feldspars have a characteristic greenish color.

Bavaria.—At Bodenmais[2] granite intersects gneisses. The ore deposits lie in cordierite gneiss. The ores contain pyrrhotite and pyrite, with some zinc blende rich in cadmium and galena rich in silver; the bodies lie in general parallel to the dip of the gneiss, but, according to Weinschenk, the contact between ore

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[1] *Zeitschr. prakt. Geol.*, vol. 16, 1908, pp. 119–122.

[2] E. Weinschenk, *Zeitschr. prakt. Geol.*, 1900, pp. 65–71.

and gneiss is sharp, though there are some disseminated sulphides in the surrounding rock. Many of the gangue minerals in the ore are rounded or corroded.

Sweden.—The renowned copper deposit at Falun,[1] in Sweden, forms a huge inverted cone enclosed in gray quartzose and gneissoid rocks and extending to a depth of 1,200 feet. The orebody is really composed of the same rock, impregnated to greater or less extent with pyrite, pyrrhotite, and chalcopyrite. The gangue minerals accompanying the ore are cordierite, magnetite, andalusite, spinel, and garnet. It is difficult to arrive at a definite conclusion regarding the origin of this deposit; at any rate it was formed at high temperature. According to Vogt the total production of copper from Falun from 1300 to the present time is about 480,000 metric tons.

The copper deposit at Bersbo,[2] in Sweden, is also considered by Bergeat to belong to this class. The ores are quartzose and are embedded in gray fine-grained "granulite" or "leptite" (p. 757), which is now by many considered an igneous and intrusive rock. In thin section the ores show a texture resembling that of contact-metamorphosed schist and contain as gangue minerals quartz, cordierite, spinel, biotite, hornblende, and garnet. On the whole, the succession is magnetite (oldest), pyrite, pyrrhotite, zinc blende, and chalcopyrite.

Norway.—A number of remarkable pyritic deposits are found in Norway; among them are the well-known localities of Röràs, Vigsnäs, and Sulitjelma, all of which have been the subject of extended discussion. They occur in metamorphic schists including clay slate, chloritic schist, amphibolite, and in part certainly in dynamo-metamorphosed gabbro intrusions. The ores consist of early pyrite, chalcopyrite, sphalerite and pyrrhotite. A little biotite and magnetite is present.[3] Flat ore lenses prevail, in some places strictly parallel to the schistosity, in other places, as at Röràs, distinctly cutting across it. In large part they are massive pyritic bodies, but the neighboring rock is usually im-

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[1] Hj. Sjögren, The Falun copper mine, *Guide exc.*, XI^e Cong. géol. internat., Stockholm, No. 31, 1910.

A. E. Törnebohm, Geol. För., *Förh.*, Stockholm, vol. 15, 1893, pp. 609–690.

[2] A. E. Törnebohm, Geol. För. *Förh.*, vol. 7, 1885, pp. 562–597.

A. Bergeat, *op. cit.*, p. 978.

[3] H. Ries and R. E. Somers, *Trans.*, Am. Inst. Min. Eng., vol. 58, 1918, pp. 244–264.

pregnated with pyritic ore. One of the flat ore-bodies at Röràs extended along its dip for 1,900 meters and was 100 meters wide, averaging 8 meters in thickness.[1] At Sulitjelma the contact phenomena have been interpreted as injections. The ore brecciates the schist and enters into it on veins and seams. Feldspar, chlorite and garnet of the schist are embayed but not sericitized or otherwise altered.

Quartz, actinolite, garnet, epidote, and biotite accompany the ore minerals at some places. Th. Kjerulf and J. H. L. Vogt[2] among others consider these deposits as igneous injections, the latter author placing them in genetic association with the gabbro intrusions.

[1] Stelzner and Bergeat, vol. 1, 1904, p. 298.

[2] Vogt has asserted a relationship between these ores and those of Rammelsberg, in the Harz, Germany. As shown on pp. 644–648, the latter belong in an entirely different class, with barite gangue, and show an absence of high temperature minerals.

CHAPTER XXX

METAMORPHOSED DEPOSITS

PROCESSES INVOLVED

Mineral deposits are usually formed during comparatively brief epochs, in which uniform conditions prevail, rendering a given set of minerals stable. In the development of the epigenetic deposits this is not invariably true, for we sometimes find evidence of successive changes in the mineral-bearing solutions; early minerals are dissolved and a new set formed. The replacement of calcite veins by silica offers an instance of this process, as do also the successive generations of minerals in zeolitic copper deposits and in pegmatite dikes.

After the epoch of mineralization has passed the deposit will, as a rule, be subjected to different temperatures and different degrees of pressure, and solutions of various kinds will percolate through it. Consequently, in many deposits the minerals of their ore are now unstable and only the slowness of the changes may prevent them from being wholly altered. "Persistent" minerals remain unaffected except by mechanical deformation, but very few minerals are persistent in all zones.

In general, when by erosion, intrusion, or dynamo-metamorphism a mineral deposit is transferred to a new zone, the characteristic minerals of this zone will develop in it and become superimposed upon the original minerals. Some deposits have a complicated history, having been subjected to several changes, each of which has left its imprint on the ores.

It often happens that a deposit becomes involved in folding or dynamic metamorphism of general or local kind; there will then be mechanical deformation; veins and irregular masses will be squeezed out into lenses which may in places overlap or imbricate. The minerals of dynamic metamorphism, such as garnet, amphibole, and biotite of the deeper zones, or chlorite, epidote, zoisite, muscovite, albite, and talc of the upper zones, will be formed from the old constituents. Hydrates may lose their water and carbonates their carbon dioxide. The quartz-sulphide veins are least affected, their minerals being comparatively persistent.

Most deposits have been exposed to static metamorphism at moderate temperature, during which chlorite, carbonates, and epidote have developed. Increased temperature may leave some deposits unaltered, while others in the vicinity of igneous masses may be profoundly modified. Examples are known of sedimentary deposits of limonite or siderite which, close to intrusive rocks, change to magnetite and specularite and in which garnets and other silicates develop. Such deposits may simulate those of contact-metamorphic origin, but in the latter the ores did not exist in the sedimentary rocks but were introduced by solutions. Some of the deposits in the pre-Cambrian terranes owe their complex nature to successive changes, and their history may be most difficult to unravel.

DEFORMED PYRITIC DEPOSITS

The copper deposit at Rammelsberg,[1] in the Harz Mountains (p. 644), illustrates well the effects of local dynamo-metamorphism at no great depth. Under strong pressure the softer minerals like galena, chalcopyrite, and zinc blende are easily deformed and pressed out to plastic streaky masses. Pyrite,[2] being harder, is crushed without plastic deformation and subsequently cemented.

W. H. Emmons[3] has examined some copper deposits in New Hampshire, the origin of which antedated the metamorphism of the surrounding rocks. At the Milan mine there are two overlapping lenses of cupriferous pyrite that are clearly portions of a single ore-body which was separated during the process of regional metamorphism. The surrounding quartz-chlorite schist was in its zone of flow while the pyrite deposit was in its zone of fracture (Fig. 268). The massive pyrite shows little deformation, but near the walls, in the lower grade ore, quartz and pyrite have been pressed out into schistose form. Thin sections show crushing and re-cementing of the pyrite, which seems massive in the hand specimen. The gangue at the Milan mine consists of quartz, muscovite, biotite, and chlorite;

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[1] *Econ. Geol.*, vol. 6, 1911, pp. 303–313.

[2] F. D. Adams, An experimental investigation into the action of differential pressure, etc., *Jour. Geology*, vol. 18, 1910, pp. 480–535.

[3] *Bull.* 432, U. S. Geol. Survey, 1910, p. 62.

probably the three last named minerals are of dynamometamorphic origin.

In the interpretation of these deposits it is necessary to search for relics of older gangue minerals, more or less affected and changed by pressure. In many deposits such minerals may have been entirely obliterated.

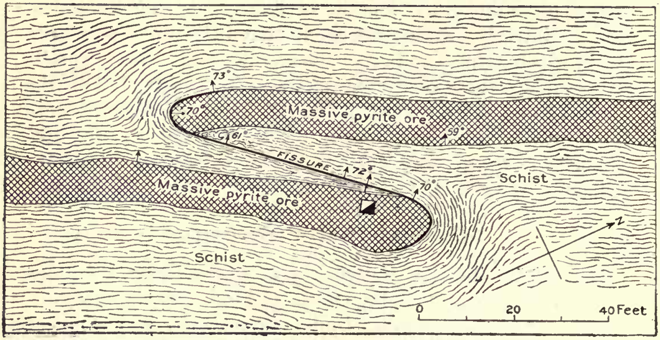

Fig. 268.—Diagram showing deformation of pyritic vein at Milan, New Hampshire. *After W. H. Emmons, U. S. Geol. Survey.*

REGIONALLY METAMORPHOSED IRON ORES

General Features.—In regionally metamorphosed sediments or in crystalline schists, the origin of which may be in doubt, bedded deposits of magnetite or specularite, or both, are often encountered. The well-known fact that iron ores such as limonite, siderite, hematite, or iron silicates (chamosite and thuringite) form integral parts of sedimentary series of all ages suggests strongly that the beds of these ores in metamorphosed rocks also had a sedimentary origin. As a rule this is no doubt true, but the metamorphism may have gone so far that the original sedimentary nature of the surrounding rocks may be open to doubt, and many observers maintain an igneous origin for some such deposits. Indirectly, igneous rocks have often brought about the accumulation of bedded iron ores, either by the weathering and denudation of intrusive rocks or lavas rich in iron, or possibly by direct emanations from volcanic rocks.

Bedded metamorphic iron ores are accompanied by silicate minerals, like feldspar, actinolite, and garnet, usually also by quartz, and they have assumed a thoroughly crystalline texture similar to that of other crystalline schists, the constituents being generally interpenetrating, indicating almost simultaneous development. Relic structure showing the sedimentary origin is rarely observed.

Swedish "Dry Ores."[1]—Sweden and Norway are rich in these bedded ores, which often appear in the vicinity of other iron deposits of different kind. Some are found near the great magmatic deposit of Kiruna, interbedded in tuff and shales of late pre-Cambrian age. Others, which are worked more extensively, appear near the metasomatic magnetites of central Sweden (p. 755) and form part of the complicated leptite series (p. 757). They are designated "dry ores" (torr-sten) and are usually siliceous, the accompanying beds averaging 84 per cent. silica. The ores average 50 per cent. iron, contained in micaceous fine-grained specularite with a little magnetite. The accompanying beds in places contain garnet, amphibole, or epidote, each mineral often forming a separate streak. They are markedly banded. Many of the beds are 10 or 15 feet thick, though some considerably exceed 15 feet, and have been followed with regular steep dip to a depth of several hundred feet. These ores contain little phosphorus. An analysis of such ore from Striberg is as follows:

Fe	52.20	CaO	1.05
Fe_2O_3	60.21	Al_2O_3	0.89
FeO	13.93	SiO_2	23.61
MnO	0.09	P_2O_5	0.043
Mg	0.31	S	0.021

Until recently little doubt has been expressed about the sedimentary origin of these ores. Lately, however, H. Johannson has announced his opinion that the fine-grained leptites are simply a product of extreme magmatic differentiation and that the accompanying bedded iron ores are also of magmatic origin. He even believes that the metasomatic limestone and "skarn ores" (p. 757) have this origin. Hj. Sjögren does not share this

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[1] H. E. Johannson, Geol. För., *Förh.*, vol. 32, 1910.
H. Sjögren, *Trans.*, Am. Inst. Min. Eng., vol. 38, 1908, pp. 766–835.
See also references on p. 755.

opinion but holds that the bedded ores and limestone ores are caused by injection or replacement by "granitic extracts" while the differentiated granulites were in the anamorphic zone. Holmquist thinks that by deep burial these originally sedimentary ores have been subjected to igneous metamorphism followed by slight regional metamorphism (p. 759).

It does not seem that the opponents to the sedimentary genesis of the ores have proved their case.

Norwegian Ores.[1]— Northern Norway is rich in deposits of the type here discussed. Banded magnetites, variously interpreted, occur on a large scale in thick beds that are traceable for several miles in South Varanger, near the Russian frontier. The ores are mined on a large scale and concentrated. Some of the larger bodies are 1,000 feet long and 25 feet thick and contain about 35 per cent. iron. One hundred million metric tons are available

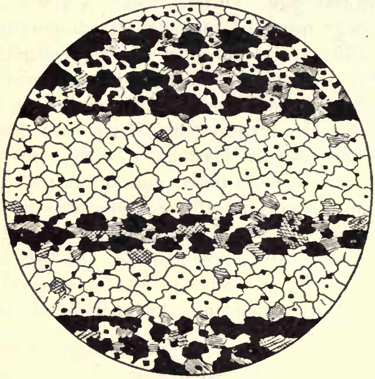

FIG. 269.—Thin section of typical Syd Varanger ore. Black, magnetite; white, quartz; striated, hornblende. Magnified 25 diameters. *After J. H. L. Vogt.*

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[1] J. H. L. Vogt, Norway, in Iron-ore resources of the world, Stockholm, 1910.

J. H. L. Vogt, Norges jernmalmforekomster, *Norges Geol. Undersök.*, No. 51, Kristiania, 1910.

Hj. Sjögren, Om jernmalmerna i granit pă Lofoten, Geol. För., *Förh.*, vol. 30, 1908.

Per Geijer, Contributions to the geology of the Sydvaranger iron-ore deposits, Geol. För., *Förh.*, vol. 33, 1911, pp. 312–343.

for open-cut mining. An analysis given by Vogt shows 36.71 per cent. Fe_2O_3, 15.40 per cent. FeO, 43.92 per cent. SiO_2, 0.07 per cent. P_2O_5, and 0.04 per cent. S. There is little alumina, lime, or magnesia. The ores are beautifully banded and according to P. Geijer are associated with fine-grained "leptites" (granulites) rich in quartz with some orthoclase and oligoclase; hornblende, garnet, and diopside accompany the ore (Fig. 269). While Vogt considers the ores to be due to igneous differentiation and Sjögren believes similar ores from the Lofoten Islands to be intrusive into an igneous rock, Geijer gives good reasons why they should be held to be of sedimentary origin and deposited as chemical sediments. It seems that the advocates of intrusive origin for these occurrences have few cogent arguments.

Ores of distinctly sedimentary origin are found at Dunderland and Naeverhaugen, also in northern Norway. They form beds traceable for many miles, with a thickness of 3 to 10 meters, or in places even 50 meters. They are intercalated in a thick series of mica schist and crystalline marbles believed to be of Paleozoic age. The closely banded ores carry mainly specularite and magnetite, but are of low grade. The concentration, attempted on a large scale and at great expense, has failed because of the scaly character of the specularite. The average content in iron is said to be 40 per cent. Hornblende, garnet, epidote, and feldspar are accessory minerals. There is little sulphur, but phosphorus is present in quantities as great as 0.3 per cent.

United States.—The Lake Superior ores do not strictly belong to the type here discussed; they are rather rich concentrations in the zone of oxidation from low-grade sedimentary beds. In places where the iron formations have been exposed to contact-metamorphism ores with magnetite and grunerite result.

Typical quartz-magnetite ores, described by Sidney Paige,[1] occur in the Llano region in Texas, but have not yet been utilized. The pre-Cambrian series of crystalline schists in this region consists of gneiss, mica schist, and quartzite with some limestone lenses. The ores are thinly bedded and occur in granular schists or gneisses. A specimen of lean ore consisted of magnetite 22 per cent., quartz 50 per cent., albite and albite-oligoclase 26 per cent. There is much more soda than potash. The iron was probably, according to Paige, deposited as glauconite, and contact-metamorphism by later pre-Cambrian granite has

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[1] *Bull.* 450, U. S. Geol. Survey, 1911.

effected the removal of potash and introduction of soda. Examples of adinole and other contact-metamorphic rocks are cited to support this view. Low-grade ore representing a bed 17 feet thick contained Fe 35.87 per cent., SiO_2 34.57 per cent., Mn 1.05 per cent., P 0.07 per cent., S 0.04 per cent., and TiO_2 0.15 per cent.

The Grenville series of pre-Cambrian metamorphosed sediments in northern New York contain, according to D. H. Newland,[1] similar deposits of magnetite. The ores are mostly enclosed in quartzose gneisses with hornblende and biotite and are believed to be of sedimentary origin.

THE ZINC ORES OF ÅMMEBERG, SWEDEN

One of the most mysterious of ore deposits is that of Åmmeberg,[2] in Sweden, where the zinc ore is disseminated in banded and contorted gray gneissoid "leptite." Some layers of gabbro, crystalline limestone, and lime-silicate rocks are intercalated in the steeply dipping leptite. Zinc blende, with very little galena, is widely disseminated in the leptite, seemingly taking the place of magnetite, and along certain zones has accumulated as long, lenticular folded bands, some of which are 30 to 50 feet in width and have been followed to depths of 1,000 feet. The ores are rich in zinc blende but contain few other minerals. A "fahlband" of disseminated pyrrhotite and arsenopyrite lies in the leptite of the footwall.

The deposit certainly seems to be of syngenetic origin and the mineral association indicates that it has been subjected to high temperatures.

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[1] Geology of the Adirondack magnetic iron ores, *Bull.* 119, N. Y. State Mus., 1908, pp. 27, 40–41.

[2] H. E. Johansson, The Åmmeberg zinc ore field, Geol. För., *Förh.*, vol. 32, Stockholm, April, 1910. Guide exc., XI^e Cong. géol. internat., Stockholm, No. 35, 1910.

CHAPTER XXXI

OXIDATION OF METALLIC ORES[1]

GENERAL CONDITIONS

The upper part of a mineral deposit, within the zone of weathering, is usually more or less altered by surface waters containing free oxygen. The direct effects of this weathering cease in many deposits at the permanent water level, but in deposits of sulphides the indirect effects, due to the action of sulphates generated by the oxidation of primary sulphides, may persist to a considerable depth below the water level. Generally speaking, the zone above the water level is that of the oxy-salts, haloid salts, and native metals; underneath this in many deposits lies a zone of varying depth in which secondary sulphides appear and strong enrichment has taken place. Finally, beneath these zones of extensive changes and molecular rearrangements is found the original or "primary" ore.

The oxidation of mineral deposits is naturally a process analogous to rock-weathering. In deposits free from sulphides the changes are relatively simple, consisting of disintegration, solution, oxidation, and hydration. Siderite alters to limonite, carbonates of manganese to pyrolusite; calcite is dissolved; the rock minerals change to kaolin. The final products are likely to be residual quartz, limonite, pyrolusite, and kaolin. Where native copper is present malachite and cuprite are also found if the leaching has not been carried too far.

In deposits which contain sulphides, but no pyrite, the changes are rather slow and inconspicuous. Galena changes slowly to anglesite and cerussite, zinc blende to calamine and smithsonite; galena and enargite often remain unoxidized close to the surface. The presence of pyrite, which easily gives off one atom of sulphur, changes and complicates the whole trend of the oxidizing processes. During oxidation the various metals behave very differently and thus many separations are effected.

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[1] For a fuller treatment of this subject see W. H. Emmons, The enrichment of ore deposits, *Bull.* 625, U. S. Geol. Survey, 1917.

DEPTH OF OXIDATION

Oxidation is a relatively slow process. Some of the more conspicuous cases of deep oxidation have required long geological time. The copper deposits at Bisbee, Arizona, for example, where large bodies of oxidized ores are present, were probably attacked by oxidation in Cretaceous time. In glaciated areas, such as Canada and northern Europe, oxidation has made little progress since the ice swept away the older accumulations of weathered products, and sulphide ores are usually found close to the surface. Comparatively little effect has been produced by an exposure of several thousand years.

In non-glaciated regions provided with a liberal rainfall the ground-water level is usually less than 100 feet below the surface and the oxidation is correspondingly shallow. On the other hand, in regions with deficient rainfall the ground water may stand several hundred feet below the surface and the oxygen has had an opportunity to decompose the ores to a similar depth. At the copper mines of Butte, Montana, the sulphides are oxidized to a depth of at most 400 feet; in the silver-gold veins of Tonopah, Nevada, 700 feet; at Tintic, Utah, in limestone, as much as 1,600 or 2,200 feet. At Bisbee, Arizona, also in limestone, oxidized copper ores are found at a depth of 1,400 feet. At the Durango lead mine, Mapimi, Mexico, the ground-water level is said to stand 2,300 feet below the surface.

In a general way the depth of the water table corresponds to the depth of the oxidized sulphides, but this is a rule with exceptions and qualifications. In most districts sulphide ores may be found in local masses almost or quite at the surface, and, on the other hand, oxidation may penetrate to a depth of several hundred feet below the water level. In the Cripple Creek district, Colorado, for example, at the Golden Cycle mine oxidized ores were found 200 feet below the water level. It is simply a question of the presence or absence of decided local movement of the oxygen-charged water downward along the fissures. The changes are, of course, greatest along the fissures, where oxidation is usually far in advance of the weathering of the general country rock. Changes in climate or elevation with corresponding changes in the water level must not be overlooked. In Arizona we find at many places—Clifton, Globe, and Ray, for instance—zones of secondary chalcocite which assuredly were formed below the water level, but which now lie high above the permanent water. At Butte, Montana, on the other hand, there is evidence of a depression of the block containing the deposits which has had the effect of raising the water level high above the position it occupied when the chalcocite enrichment took place. The facts observed in some districts can be explained only on the supposition of repeated and relatively rapid changes of water level.

The temperature also plays a part. We may expect a deeper oxidation in warm climates than in cold climates simply because heat accelerates reactions. Frequent alternation of moisture and dryness promotes oxidation. Porosity and fissuring of the rocks and ores are factors extremely favorable to oxidation.[1]

In a region of dry climate where mountain ranges are separated by valleys filled with saline deposits, the winds carry the salt to the oxidizing outcrops and the development of chloride of silver, for instance, is facilitated.[2]

The essential factors entering into the problem of oxidation of ore deposits are, then, ore; metal; country rock; fissuring; permeability; climate, water level, and rainfall; topography; geological age and history of deposit.

OUTCROPS

The outcrops of deposits in glaciated areas are likely to be inconspicuous, except where the principal gangue mineral is unusually hard, like quartz. In non-glaciated regions the outcrop form is determined by the difference in the rate of erosion of the deposit and the country rock.[3] A thick and hard quartz vein or a mass of solid silicified rock or garnet rocks in contact-metamorphic deposits will remain as little ridges or series of knobs above the general level of a softer country rock. The quartz veins of California, for instance, are ordinarily easily traceable on the surface. Where the quartz contains much pyrite, a honeycombed or cellular mass of limonite and quartz remains more or less conspicuously above the surrounding

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[1] A. M. Finlayson, Economics of secondary enrichment, *Min. and Sci. Press*, July 16 and 23, 1910.

[2] C. R. Keyes, Origin of certain bonanza silver ores of the arid region, *Trans.* Am. Inst. Min. Eng., vol. 42, 1911, pp. 500–517.

[3] W. H. Emmons, Outcrop of ore-bodies, *Min. and Sci. Press*, Dec. 4 and 11, 1909.

country rock. Such weathered croppings the German miner calls "eiserner Hut," the Cornishman a "gossan," the Australian "ironstone." The Spanish-American gives these oxidized limonite ores the names "colorados," "pacos," "podridos," or "quemados," according to their reddish color, their soft or rotten character, or their burnt appearance.

Where the minerals of the deposit are softer than the country rock a depression, or little gap, or saddle may mark its outcrop. At Butte the outcrops of the rich copper veins, which contain little gangue, are inconspicuous, while the silver veins can be easily followed along the surface. Along a single vein there may be great variation in the croppings. Barren parts tend to stand up prominently, while the ore shoots, containing softer metallic minerals, may easily become effaced at the surface.

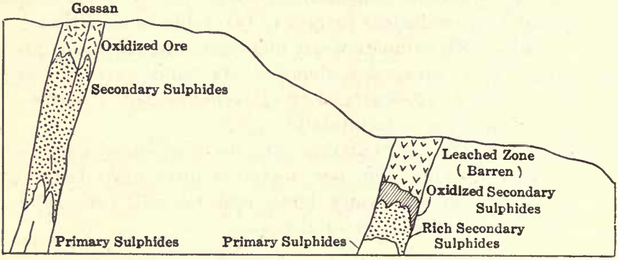

FIG. 270.—Diagram showing normal course of oxidation in pyritic veins and influence of rapid erosion on exposed secondary sulphide zone. In the deposit to the right the gossan has been eroded and the upper part of the secondary sulphide zone leached, causing a thinner but richer secondary zone.

The typical gossan of pyritic bodies, under uniform conditions of high water level and slow erosion, probably remains without much change for long periods. When a gradual lowering of the water level and a quickening of the erosion expose new parts of the deposit to the decomposing influence of oxygenated waters the transfer of material downward becomes more active, and in a copper deposit, it may happen that the surface portion becomes entirely leached of metallic minerals and consists simply of cellular quartz and of the more resistant parts of the country rock. Some such croppings of pyritic copper ores contain scarcely a trace of iron or copper (Fig. 270) (p. 858).

NOMENCLATURE

The terms primary and secondary as applied to ores are inconvenient and often misleading. They will, therefore, be avoided as far as possible.

A great number of ore deposits are formed by ascending waters. Such waters and such deposits are termed *hypogene*.[1] Most changes during direct and indirect oxidation are caused by descending surface waters. Such waters and the ores formed by them by the rearrangements of the hypogene ores are called *supergene*.[1]

In many cases, valueless but mineralized material has been worked over by descending surface waters and valuable ore-bodies have been formed from it. Ransome has proposed the term *protore*[2] to designate any primary material of too low tenor to constitute ore but which may be concentrated into ore. Thus, the low grade pyritic dissemination underneath the chalcocite blanket at Ely, Nevada, is protore.

PRINCIPLES OF OXIDATION

The powerful reagents of oxidation are oxygen, carbon dioxide, and sulphuric acid. The last combines with iron to form ferric and ferrous sulphate, the former being an especially active agent of oxidation, while the latter is an important reducing agent. Sodium chloride and sulphuric acid yield hydrochloric acid, which easily combines with iron to make the strongly oxidizing ferric chloride. Under the influence of sulphuric acid the waters change from the calcium carbonate type characteristic of the normal surface conditions to the calcium sulphate type. The aluminous silicates are attacked by sulphuric acid and by carbon dioxide; sulphates, carbonates and hydrous silicates result.

Insoluble minerals, like cassiterite, wolframite, and often also gold, remain without change in the outcrops, enrich them upon contraction of volume, or on their disintegration are concentrated into placers. Soluble salts, especially sulphates, are carried away. Newly formed compounds are precipitated, chiefly by reactions between carbonates and sulphates or between sulphates. Below a certain point, usually the water

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[1] F. L. Ransome, *Bull.* 540, U. S. Geol. Survey, 1912, p. 152.

[2] W. H. Emmons, *Bull.* 625, U. S. Geol. Survey, 1917, p. 68.

level, the free oxygen rapidly diminishes and sulphides are precipitated by reactions between sulphates and sulphides or by other processes.

Much of the dissolved material is naturally removed by the running water of the vicinity, but the greater part of it sinks in the deposit itself and there becomes re-deposited, thus contributing to the general process of enrichment by the descending waters. Some enriched deposits are the product of long-continued descending concentration from a great thickness of rocks now removed by erosion.

In ore deposits free from great amounts of resistant quartz gangue oxidation obliterates structure. Heavy pyritic deposits appear at the surface as cellular masses of quartz and limonite; the sheeted gold-telluride veins of Cripple Creek, Colorado, which do not carry much quartz, appear as brown clayey bands without visible structure. There is thorough rearrangement of metal association, and often also segregation of new minerals in large masses. Limestone country rock especially favors such changes. Lead and zinc closely associated in galena and zinc blende part company; the oxidized zinc ores wander farther away from the original deposit than does the cerussite. Copper and iron, so intimate in primary ores, separate more or less in the zone of oxidation, the former exhibiting a centripetal, the latter a centrifugal tendency,[1] and arrange themselves concentrically, just as happens in fragments of sulphide ore subjected to "kernel roasting."

Masses of nearly pure kaolin and alunite often form in this zone.

Some sulphates, like anglesite or basic ferric sulphate, are insoluble: others, like goslarite (Zn), mallardite (Mn), epsomite (Mg), ferrous sulphate, and aluminous sulphate, are most easily carried away. Gypsum, common as a product of exchange in reactions leading to the formation of insoluble carbonates in limestone, is also rather easily removed in solution. The native carbonates of zinc and copper are relatively insoluble and may remain for a long time in the gossan. Other minerals characteristic of the oxidized zone are native metals (copper, gold, silver, and mercury), chloride, bromide, and iodide of silver, phosphates, arsenates, antimoniates, molybdates, vanadates, rarely chro-

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[1] W. Lindgren, L. C. Graton, and C. H. Gordon, The ore deposits of New Mexico, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 55; see also Pl. IV, *B.*

mates; also hydroxides and oxychlorides; and a few hydrous silicates, like calamine and chrysocolla.

There is then, during oxidation, both dissipation and concentration of metals. The concentration may take place either in the deposit itself or may be effected by the precipitating influence of the country rock on the solutions; bodies of oxidized zinc ores often form in the limestone surrounding a deposit.

In the zone of supergene sulphides below the direct oxidation we meet the copper sulphides—mainly chalcocite and covellite, rarely chalcopyrite and bornite; also, argentite, and complex sulphantimonides and sulpharsenides, associated with native silver. Pyrite and zinc blende are seldom found as products of this zone.

Generally speaking, solution prevails near the surface and precipitation and cementation[1] in the supergene sulphide zone. In the zone of direct oxidation enrichment may or may not take place. If there is a supergene sulphide zone this always involves enrichment.

The character of the solutions changes gradually in depth. Oxygen is removed; the free acid decreases; reduction replaces oxidation; gases like H_2S and CO_2 may be generated. The general character of gangue and wall rock is of great importance. If carbonates prevail, the minerals that form may differ from those that are developed in a quartzose gangue. The results show an infinity of variations and complexity.

TEXTURES AND CRITERIA OF THE OXIDIZED ZONE

The action of meteoric waters, aided by sulphuric acid in pyritic deposits, opens spaces in the zones of oxidation resulting in vuggy, cellular, honeycombed structures with clayey masses if silicates are present. Deposition proceeds in part by replacement without constancy of volume, in part by crustification in open cavities; mammillary, concretionary, and stalactitic forms are common, alternating with crusts of crystallized minerals and pseudomorphs stable at the particular moment. Nodular textures are common coupled with rearrangement of the oxy-salts in shells, so that copper ores may surround limonite or zinc ores have a core of cerussite. Reticulating fractures are filled with oxidized products. Concentric rings of the same products

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[1] P. Krusch, Die Eintheilung der Erze, etc., *Zeitschr. prakt. Geol.*, 1907, pp. 129–139.

surround sulphides (Fig. 271). Colloidal solutions and suspensions are equally common as electrolytes. Solutions change rapidly in composition so that calcite and even quartz crusts may alternate with limonite, kaolin and oxidized salts of copper, lead and zinc. The volume is diminished and enrichment of relatively insoluble constituents follows.

The presence of limonite with other hydrous oxy-salts is one of the safest criteria of oxidation, but the absence of sulphides is not necessary for oxidation proceeds extremely capriciously and residual sulphides may be found at all levels in the zone of oxidation. To a limited extent supergene sulphides like chalcocite, covellite and argentite may be formed in the oxidized zone wherever there was a temporary shortage of oxygen.[1]

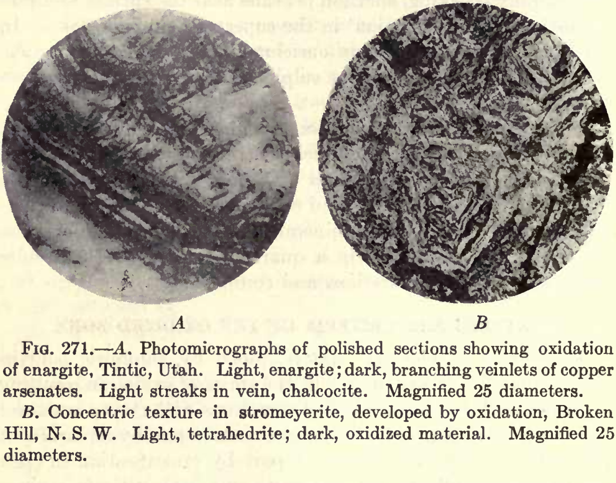

Fig. 271.—*A*. Photomicrographs of polished sections showing oxidation of enargite, Tintic, Utah. Light, enargite; dark, branching veinlets of copper arsenates. Light streaks in vein, chalcocite. Magnified 25 diameters.

B. Concentric texture in stromeyerite, developed by oxidation, Broken Hill, N. S. W. Light, tetrahedrite; dark, oxidized material. Magnified 25 diameters.

TEXTURES OF THE SUPERGENE SULPHIDE ZONE

In the zone of supergene sulphides, important in copper and silver deposits, precipitation and deposition prevails and the textures become more compact, though in places loose, earthy ores are present. The so-called "sooty" chalcocite at Butte and

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[1] W. Lindgren, *Econ. Geol.*, vol. 10, 1915, p. 236.

W. Lindgren and G. F. Loughlin, *Prof. Paper* 107, U. S. Geol. Survey, 1919.

many other places exemplifies this latter condition. The supergene sulphide ores are permeable and more or less porous. With increasing compactness replacement becomes more evident and proceeds volume for volume. The secondary sulphides replace the primary ore minerals in manifold succession and form. Reticulating veinlets are mostly formed by replacement; grains and crystals of pyrite are concentrically replaced by chalcocite (Figs. 275, 276). The supergene sulphides also fill veinlets or vugs or form thin coatings. Regular "graphic intergrowths" often similar to eutectic textures in metals develop at many places (Fig. 272). This was formerly thought a criterion of primary contemporaneous origin but it is now known to be a feature of replacement most commonly of supergene origin.[1] Thus chalcocite replaces bornite and covellite replaces galena;[2] stromeyer-

FIG. 272.—Intergrowth of bornite (*b*) and chalcocite (*cc*). Wall mine, Virgilina, North Carolina. *cc* (*sec*), Secondary chalcocite. Magnified 50 diameters. *After L. C. Graton and J. Murdoch.*

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[1] F. B. Laney, *Econ. Geol.*, vol. 6, 1911, p. 399.

C. C. Gilbert and G. E. Pogue, *Proc.* Nat. Mus., vol. 45, 1913, pp. 609–625.

L. C. Graton and J. Murdoch, *Trans.*, Am. Inst. Min. Eng., vol. 45, 1913, p. 768.

J. Segall, *Econ. Geol.*, vol. 10, 1915, p. 469.

W. L. Whitehead, *idem*, vol. 11, 1916, pp. 1–13.

F. N. Guild, *idem*, vol. 12, 1917, pp. 297–353.

[2] W. L. Whitehead, *op. cit.*

ite replaces chalcopyrite and galena; proustite replaces galena. Generally speaking, the supergene sulphides replace primary sulphides but very rarely primary gangue minerals.

SOLUTION

In the presence of water oxygen attacks the sulphides and carbon dioxide the silicates. Alkaline solutions would attack the quartz and the silicates but under the influence of free sulphuric acid generated by oxidation of pyrite they are generally absent. Distinction must be drawn between solution and decomposition; most of the changes in the oxidized zone involve decomposition. In general oxidation tends to transform sulphides, sulphosalts, arsenides, tellurides, etc., to oxygen salts and native metals both of which may be further changed or carried away by surface waters. The silicates in the deposit are changed to a few stable minerals: kaolin, limonite, manganese dioxide and quartz. Carbonates of earthy metals and alkalies are carried away; original quartz is rarely attacked but new silica from the decomposition of the silicates may be deposited as opal and chalcedony.

Oxidation tends to thorough change of composition and to obliteration of original texture and structure.

The simple sulphides are very slightly soluble in water, the solubility decreasing as follows: Mn, Fe, Ni, Cd, Zn, Cu, Pb, As, Ag, Bu, Hg.[1]

In dilute sulphuric acid (1/10 normal), pyrrhotite, chalcopyrite, galena, zinc blende and cadmium sulphide are dissolved or readily attacked. Argentite, galena, bornite, arsenopyrite, stibnite, pyrargyrite and polybasite are slightly attacked,[2] while many others like cinnabar, molybdenite, realgar, orpiment, bismuthinite, covellite, and chalcocite are not attacked.

The decomposition or solution is often hastened if the dilute sulphuric acid contains ferric sulphate. Few sulphides resist this reagent.[3]

Some sulphides react with alkaline solutions at ordinary tem-

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[1] Oscar Weigel, *Zeitschr. physikal Chemie*, vol. 58, 1907, pp. 293–300.

[2] H. C. Cooke, *Jour. Geology*, vol. 21, 1913, pp. 1–28.

[3] For experimental data see G. S. Nishihara, The rate of reduction of acidity of descending waters by certain ore and gangue minerals, *Econ. Geol.*, vol. 9, 1914, pp. 743–757.

perature.[1] Orpiment, realgar, stibnite and pyrrhotite are strongly attacked by a 1 per cent. solution of $NaHCO_3$; many others are slightly attacked; arsenopyrite, cinnabar, enargite, chalcocite, bornite, light colored zinc blende and niccolite are practically resistant.

Experiments with pyrite have not given consistent results. A. N. Winchell[2] exposed powdered pyrite for 10 months to the action of distilled aerated water and obtained a very slow rate of oxidation, the solution containing $Fe_2(SO_4)_3$ and H_2SO_4. His results were in general confirmed by F. F. Grout.[3] H. A. Buehler and V. H. Gottschalk[4] obtained a much more rapid attack and in 3 months the filtrate yielded 2.5 to 3.7 per cent. of the original weight of the iron in the sample. Sphalerite in the same time yielded only 0.2 per cent. of its zinc, galena 0.005 per cent. of its lead, covellite 2.7 per cent. of its copper, and chalcopyrite 1 per cent. of the same metal. On the other hand, enargite showed no solubility. When the various sulphides were mixed with pyrite the action was much more energetic. In the time specified sphalerite with pyrite yielded 4.2 per cent. of its zinc, galena with pyrite 0.7 per cent. of its lead, covellite with pyrite 2.7 per cent. of its copper, covellite with marcasite 27.6 per cent., and enargite with pyrite 10 per cent. of its copper. After an exposure of only 7 weeks, pyrite had oxidized to the amount of 0.1 to 0.28 per cent. of its original weight.

In a second paper Gottschalk and Buehler[5] show that while in a mixture of two sulphides there is a large increase in the solution of one, there is also a protective action exerted on the other; and further that there exists a difference of potential between the sulphides, which can be arranged in a series similar to the electrolytic series of metals. Acceleration of reaction is due to electric currents generated by contact of minerals of different potential; the currents flow from the mineral of the higher potential, and the mineral of lower potential will dissolve more rapidly. In mixtures with pyrite the iron transferred is

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[1] F. F. Grout, On the behavior of acid sulphate solutions of copper, silver and gold with alkaline extracts of metallic sulphides, *Econ. Geol.*, vol. 8, 1913, p. 427.

[2] A. N. Winchell, *Econ. Geol.*, vol. 2, 1907, pp. 290–294.

[3] *Idem*, vol. 3, 1908, p. 532.

[4] *Idem*, vol. 5, 1910, pp. 28–35.

[5] *Econ. Geol.*, vol. 7, 1912, pp. 15–34.

but a small portion of that obtained when iron disulphide is treated alone.

From this it follows that the order of oxidation in a mixture of minerals varies with conditions of mass, aggregate and character of solution. No general rule can be formulated. It is known, however, that, for instance, zinc blende oxidizes before chalcopyrite and the latter before pyrite. In a pyrite-chalcocite ore the chalcocite is attacked first.

The relative solubility of the various carbonates and sulphates are important for the distribution of metals in the oxidized zone. The following data have been determined by Kohlrausch. The low solubilities of the carbonates are considerably increased in the presence of CO_2.

SOLUBILITY OF SULPHATES AND CARBONATES AT 18° C. IN GRAMS OF ANHYDROUS SALT PER 100 GRAMS OF H_2O.

Salt	Grams	Salt	Grams
$BaSO_4$	0.00023	$PbCO_3$	0.0001
$PbSO_4$	0.0041	$CaCO_3$	0.0013
$CaSO_4$	0.20	Ag_2CO_3	0.0017
Ag_2SO_4	0.55	$BaCO_3$	0.0023
K_2SO_4	11.11	$ZnCO_3$	0.0047
Na_2SO_4	16.83	$MgCO_3$	0.100
$CuSO_4$	19.30	$FeCO_3$	0.073[4]
$FeSO_4$	23.00	$MnCO_3$	....
$Al_2(SO_4)_3$	31.30[1]	$CuCO_3$	....
$NiSO_4$	34.20[2]	Na_2CO_3	19.38
$MgSO_4$	35.43	K_2CO_3	108.00
$ZnSO_4$	53.18		
$MnSO_4$	65.00[3]		

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[1] With 18 mol. H_2O at 0°; 89.1 with 18 mol. H_2O at 100° C.

[2] At 15°.

[3] At 30°.

[4] Sat. with CO_2 at 7 at.

The descending solutions contain many salts and in places colloids, which even in the presence of electrolytes may fail to be coagulated, if "protected" by the presence of certain other colloids. Near the surface in pyritic deposits ferric sulphates and even aluminum sulphate and free sulphuric acid may be abundant but with increasing depth the ferrous sulphate predominates and the solutions tend to become neutral. Gases like CO_2 and H_2S may be generated locally. In general, during oxidation there is a great dissipation of the sulphates of iron, zinc and calcium.

PRECIPITATION

Precipitation is effected by reactions between solutions, by hydrolysis, by coagulation, by gases and by reactions between the solutions and solids. The latter phase is very important. Many reactions take place by the action of solutions on sulphides or on gangue minerals or on country rock of sedimentary or igneous origin. The investigations of E. C. Sullivan[1] have shown that silicates may precipitate oxygen salts by chemical reactions. In a general way this was known long ago, and E. Kohler[2] showed that cupric sulphate lost its copper when filtered through kaolin. This was attributed to adsorption—that is, an accumulation of dissolved substance on the contact between liquid and solid—but Sullivan shows that a chemical change takes place. The natural silicates such as kaolin, albite, orthoclase, amphibole, pyroxene, and mica precipitate the metals from salt solutions, while at the same time the bases of the silicates are dissolved in quantities

Fig. 273.—Photomicrograph of thin section showing azurite crystals replacing kaolin, Clifton, Arizona. Magnified 15 diameters.

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[1] The interactions between minerals and water solutions, *Bull.* 312, U. S. Geol. Survey, 1907; *Econ. Geol.*, vol. 1, 1905, p. 67.

[2] *Zeitschr. prakt. Geol.*, vol. 11, 1903, p. 49.

nearly equivalent to the precipitated metals. The latter precipitates take the form of hydroxides or basic salts, though silicates may also be formed to some extent. Thus by a simple chemical exchange the metal may be removed from a solution and fixed in the solid state and thus concentrated by contact with even the most insoluble of silicates.

These experiments elucidate the deposition of brochantite and chrysocolla in granitic and porphyritic rocks, as well as the deposition of cuprite and azurite in shale.[1] (Fig. 273.) A solution of silver sulphate yielded its metal completely to a powdered clay gouge, metallic silver being probably formed. With kaolin the reaction is rapid and the copper solution soon becomes colorless. The iron in ferric and ferrous sulphate is easily retained by kaolin as limonite.

The direct oxidation of galena, for instance, yields carbonate and sulphate of lead. By further reactions with ferric sulphate basic sulphates of iron and lead are formed like plumbojarosite, and in this way the lead is further distributed. Many other difficultly soluble basic sulphates form during oxidation; alunite is one of the most common of these.

SUPERGENE SULPHIDES

The development of secondary sulphides may take place by direct precipitation from solutions by means of hydrogen sulphide or other reducing solutions or gases; or it may result from a metasomatic interchange between a solution and a solid, usually another sulphide. Dilute sulphuric acid generated by the decomposition of pyrite, for instance, attacks a few sulphides, with the evolution of hydrogen sulphide. This gas is produced in abundance by the attack on pyrrhotite and to a less extent, according to R. C. Wells, when zinc blende is exposed to the acid. If copper is present in the solutions, a precipitate of cupric sulphide (CuS) will be formed, besides some cuprous sulphide(Cu_2S). Sulphides are formed mainly where the supply of oxygen from the surface becomes nearly exhausted.

Previous to the year 1900 the presence of secondary sulphides as indirect products of oxidation had been noted by some observers and had been definitely stated by L. de Launay.[2] In the

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[1] W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905, p. 191.

[2] Les variations des filons métallifères en profondeur, *Revue générale des Sciences, etc.*, No. 8, April 30, 1900.

year referred to S. F. Emmons, C. R. Van Hise, and W. H. Weed in three notable papers[1] formulated the important law of the accumulation of sulphides as a concentration from the overlying oxidized zone, at or below the water level. It was shown that in copper deposits chalcocite and covellite were precipitated by pyrite from sulphate solutions and that under similar conditions in silver deposits argentite, stephanite, polybasite, and pyrargyrite or proustite might form; it was also shown that zinc blende and galena were probably precipitated in a similar manner. The chemical reasons for these reactions were found in the so-called Schürmann's law,[2] by which it was established that in the presence of the sulphides of certain of the metals the salts of other metals would be decomposed and the metals precipitated as sulphides. This was thought to indicate that the metals which were precipitated possessed a greater affinity for sulphur than the other metals.

Schürmann's series was as follows: Mercury, silver, copper, bismuth, cadmium, lead, zinc, nickel, cobalt, iron, and manganese. The solution of a salt of any of these metals will be decomposed by the sulphide of any succeeding metal and the first metal will be precipitated as a sulphide. Thus from a solution of silver or copper salts the metal would be precipitated by a sulphide of lead, zinc, or iron. If secondary deposition of sulphides by reaction of pyritic ores on descending sulphate waters had taken place in an ore deposit containing silver, copper, lead, and zinc, these sulphides would theoretically be arranged in the following order: Argentite, chalcocite, galena, and zinc blende, the last at the lowest level. It was shown later by R. C. Wells[3] that the influencing factor was the relative solubility of the sulphides (p. 838) rather than the "affinity for sulphur."

The farther apart any two sulphides are in Schürmann's series the more nearly complete is the replacement. The full series is not represented by natural sulphides and in ore deposits the reactions of copper and silver solutions are the most important. Supergene sulphides of bismuth, nickel and cobalt

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[1] S. F. Emmons, The secondary enrichment of ore deposits.
C. R. Van Hise, Some principles controlling the deposition of ores.
W. H. Weed, Enrichment of gold and silver veins.
All in *Trans.*, Am. Inst. Min. Eng., vol. 30, 1901.

[2] E. Schürmann, *Leibig's Ann. der Chemie*, vol. 249, 1888, pp. 326–350.

[3] *Econ. Geol.*, vol. 5, 1910, p. 14.

are not known and supergene cinnabar is rare. On the other hand chalcocite and covellite replace galena and zinc blende as well as pyrite and they may also replace sulphosalts, such as enargite and tetrahedrite and iron-copper sulphides like bornite and chalcopyrite; argentite replaces the sulphides of lead, zinc and iron; galena replaces zinc blende though the reaction is of slight economic importance. The supergene sulphosalts of silver such as polybasite replace galena and other sulphides. Complex sulphosalts of lead, *e.g.*, jamesonite may replace simple sulphides as well as galena.

Supergene sulphide deposition is accompanied by few, if any, characteristic gangue minerals; opal, chalcedony and kaolin are occasionally present; quartz is rare.

The rôle of colloidal solutions may also involve the transportation of sulphides. It has long been known that sulphides may be transferred into colloid solutions by certain dispersing agents. Experiments by John D. Clark and others[1] have shown the extent of this possibility.

Nearly all sulphides, arsenides and sulphosalts may become highly dispersed as colloids under the influence of hydrogen sulphide in solutions of mild alkalinity. These minerals are then in condition to migrate with the solutions. With the escape of H_2S or by contact with calcareous and argillaceous material precipitation and replacement may occur. Minerals may crystallize directly from colloid solutions. Secondary upward migrations may thus occur in connection with repeated invasions of hypogene solutions. Where H_2S is locally developed this process may be of some importance in the supergene sulphide zone.

A. F. Rogers[2] has suggested that enrichment by secondary chalcocite has taken place by such ascending solutions, specially citing the case of the Butte deposits. This can not be regarded as proved.

The temperature during direct oxidation of pyritic ores may in places rise considerably above 50° C. It is probable that even in the supergene sulphide zone fairly high temperatures of 30° or 40° C. may obtain at times.[3]

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[1] C. F. Tolman and John D. Clark, *Econ. Geol.*, vol. 9, 1914, pp. 559–592. John D. Clark and P. L. Menaul, *Econ. Geol.*, vol. 11, 1916, pp. 37–41.

[2] *Econ. Geol.*, vol. 8, 1913, pp. 781–794.

[3] W. H. Emmons, *Econ. Geol.*, vol. 10, 1915, pp. 151–160.

It has already been stated that in deep oxidized zones supergene sulphides may well form together with oxy-salts. Generally, however, the sulphides begin at the water level and extend for a varying distance below it thus forming a zone of secondary sulphides, which may be many hundreds of feet deep or which may only occupy a thickness of a few feet. Permeability of the primary ore has much to do with this but time and climatic conditions are also potent factors. The water level may have changed its position during geologic times and so we may find chalcocite zones "marooned" high above the water level and now in active process of oxidation. In the same way the products of direct oxidation of a former low water level may now be buried in the underground water.

In such cases physiography may come to the rescue and elucidate the changes which have taken place in the configuration of the ground.[1]

Such well-defined zones of supergene sulphides are common only in the case of copper and silver. In the case of silver deposits the products of direct oxidation and sulphide deposition are greatly mingled. No great zones of supergene lead or zinc sulphides are known.

Certain elements like iron, zinc and arsenic, which may be common in the primary ore, may be completely eliminated in the oxidized ore and in the supergene sulphides.

CRITERIA OF SUPERGENE SULPHIDE ENRICHMENT

The question whether or not secondary sulphides have been deposited in an ore-body by descending waters is most important. If the ore minerals are only a part of a shallow enriched layer and poorer ore is to be expected at lower levels, knowledge of this fact is greatly to be desired from the mine owner's standpoint.

The best geological evidence of enrichment consists in the progressive, uniform impoverishment of all similar sulphide deposits in a given district, coupled with the condition that the change in ore should be dependent upon post-mineral topographic development.[2] If the enriched zone is shallow such evidence may be conclusive. If it is deep there may be difficulties in arriving at a correct conclusion.

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[1] W. W. Atwood, The physiographic condition at Butte, Mont., *Econ. Geol.*, vol. 11, 1916, pp. 697–740.

[2] F. L. Ransome, *Econ. Geol.*, vol. 5, 1910, pp. 205–220.

The occurrence of exceptionally rich ores just below the zone of oxidation is generally suggestive of enrichment. Graton and Murdoch[1] find the important criterion in the microscopic structure of the ore but this is not always reliable because of the similarity of the latest hypogene replacements to those of the descending surface waters. If microscopic structure were always a safe guide, we should know more about the genesis of the great copper deposits at Butte than we now do.

Generally speaking it is believed that the presence of chalcocite and covellite in large amounts is a safe indication of supergene sulphide enrichment, while in silver deposits, the occurrence of rich silver sulphantimonides and argentite just below the zone of oxidation is also, in most cases, a reliable criterion.

IRON

In iron deposits with siderite and iron silicates oxidizing conditions result in abundant limonite. Hematite and magnetite deposits are very slowly oxidized but ultimately form some limonite. The small amount of magnetite in certain deposits of hematite and limonite may be residual or it may have been generated during mild regional or contact metamorphism. Oxidation under tropical conditions generates hematite from ferrous silicates in rocks.

It is known that magnetite may alter to hematite; pseudomorphs (martite) of hematite after magnetite are not uncommon. The nature of this alteration is as yet in some doubt. Some authors[2] noting the abundance of hematite near the surface in some magnetite deposits and the anastomosing veinlets of hematite in the magnetite itself have believed that this slight oxidation of a resistant mineral is caused by descending surface waters. Undoubtedly the two minerals may form together independently of oxidation, or hematite may be introduced later than the magnetite during high temperature processes.

It has been stated lately[3] that the iron oxides are in part solid

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[1] *Trans.*, Am. Inst. Min. Eng., vol. 45, 1913, pp. 26–31.

[2] W. Lindgren and C. P. Ross, The iron deposits of Daiquiri, Cuba, *Trans.*, Am. Inst. Min. Eng., vol. 53, 1916, pp. 40–66.

See also on same subject, M. Roesler, *Trans.*, Am. Inst. Min. Eng., vol. 56, 1917, pp. 77–127.

[3] R. B. Sosman and J. C. Hostetter. *Trans.*, Am. Inst. Min. Eng., vol. 58, 1918, pp. 409–444.

solutions of Fe_3O_4 in Fe_2O_3 while others are mixtures of the two compounds. Metallographic methods do not appear to have been used in this examination.

In sulphide deposits pyrite, pyrrhotite and marcasite are the principal iron minerals.[1]

Pyrite is a persistent mineral forming in all deposits and at all temperatures even locally at the surface under reducing conditions. It may be reproduced in alkaline solutions, or, with marcasite, in slightly acid solutions.

The oxidation of pyrite is started by oxygen and hastened by the ferric sulphate developed.

$$FeS_2+7O+H_2O=FeSO_4+H_2SO_4.$$

This reaction involves several intermediate stages during which sulphur dioxide, sulphur, or hydrogen sulphide may form. The well-known smell from old dumps containing pyrite indicates the development of sulphur dioxide, according to the equation $FeS_2+6O+H_2O=FeSO_4+H_2SO_3$, and this sulphurous acid is further oxidized to sulphuric acid. The presence of sulphur is often observed near the surface in the casts of dissolved pyrite crystals.

Ferrous sulphate easily changes to the ferric salt and to the hydroxide:

$$2FeSO_4+H_2SO_4+O=Fe_2(SO_4)_3+H_2O.$$

$$6FeSO_4+3O+3H_2O=2Fe_2(SO_4)_3+Fe_2(OH)_6.$$

Ferric sulphate hydrolyzes to hydroxide and free acid:

$$Fe_2(SO_4)_3+6H_2O=2Fe(OH)_3+3H_2SO_4.$$

The ferric sulphate is a strong oxidizing agent, which, according to Stokes, attacks pyrite rapidly at 100° C. and more slowly in the cold:

$$Fe_2(SO_4)_3+FeS_2=3FeSO_4+2S.$$

The sulphur may be oxidized to sulphuric acid. The colloid ferric sulphate changes easily to various basic sulphates, like coquimbite, copiapite, or jarosite, often found in the lower part of the oxidized zone. Limonite is usually the final product. Melanterite ($FeSO_4.7H_2O$) often forms as crusts and stalactites by dripping mine waters.

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[1] E. T. Allen, J. L. Crenshaw and John Johnson, The mineral sulphides of iron, *Am. Jour. Sci.*, 4th ser., vol. 33, 1912, pp. 169–236.

Ferrous sulphate and calcite yield limonite and soluble gypsum. In deposits in limestone this is a most important reaction. The ferric sulphate and the sulphuric acid may attack other sulphides present, such as chalcopyrite or zinc blende. In the deeper levels of the oxidizing zone much or all of the sulphate is likely to remain in the ferrous form. The sulphate solutions sink to the ground-water level and may here produce manifold changes by reaction with primary sulphides.

Marcasite forms only in acid solutions, but may then crystallize together with pyrite. Above 450° C. it passes into pyrite. Low temperature and free acid favor its development. Marcasite is thus a relatively unstable mineral formed mainly near the surface. A few ore deposits of igneous affiliations like that at Goldfield, Nevada, contain marcasite, but it is usually a very late product forming in fissures and vugs. It is most prominent in the lead-zinc ores of the Mississippi valley type. Nevertheless it is not found among the supergene sulphides probably because descending sulphates are likely to attack it vigorously.

It oxidizes much more easily than pyrite but the reactions are the same.

Pyrrhotite, regarded as a solid solution of sulphur in FeS, is in nature a high temperature mineral not known to occur in ores of shallow or intermediate depths. It is readily attacked by dilute H_2SO_4 with evolution of H_2S, which in copper deposits may precipitate copper sulphide and prevent the development of deep chalcocite zones. It is also easily attacked by oxidation, the H_2SO_4 formed accelerating its destruction.

$$Fe_7S_8+31O+H_2O=7FeSO_4+H_2SO_4.$$

COPPER

Minerals.—The most common primary sulphides of copper and iron include chalcopyrite ($CuFeS_2$) and bornite (Cu_5FeS_4). The pale yellow mineral chalmersite ($CuFe_2S_3$) is probably more common than has been suspected.[1] There are further tetrahedrite ($Cu_8Sb_2S_7$) with its arsenical analogon tennantite, and enargite (Cu_3AsS_4), all three important copper ores in some places. Bournonite ($CuPbSbS_3$) is sometimes an ore mineral and there are a number of rare copper and lead-copper sulph-bis-

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[1] B. L. Johnson, *Econ. Geol.*, vol. 12, 1917, pp. 519–525.

muthides. Native copper is an important primary ore mineral in some districts while the arsenides are rare minerals.

Chalcocite (Cu_2S) is rarely, if ever, of hypogene origin.

The supergene copper minerals are very numerous. Those of some economic importance include native copper, cuprite (Cu_2O), several indefinite minerals of colloid origin containing CuO, M_2O, ZnO, SiO_2 and H_2O (copper pitch ores), also the oxychloride atacamite ($Cu_2Cl(OH)_3$) which is found where windblown sodium chloride is available.

There are further chalchantite ($CuSO_4.5H_2O$) with its related minerals kröhnkite ($CuSO_4.Na_2SO_4+2H_2O$) and natrochalcite ($Na_2SO_4.Cu_4(OH)_2(SO_4)_2+2H_2O$), all important minerals in arid countries like Chile. Phosphates are rare.

Malachite ($Cu_2(OH)_2CO_3$), azurite ($Cu_3(OH)_2(CO_3)_2$) and chrysocolla ($CuSiO_3.2H_2O$) are the most abundant of all the oxidized copper minerals. Basic sulphates like brochantite ($Cu_4SO_4(OH)_6$) are locally abundant as are various arsenates most common among which is olivenite ($Cu_3As_2O_8.Cu(OH)_2$).

The secondary sulphides of prime importance are chalcocite (Cu_2S) and covellite (CuS) while bornite and chalcopyrite may also be of supergene origin.

Solutions and Precipitation.—Copper is one of the most easily transported metals and as it is also easily precipitated its supergene deposits have great importance. Copper migrates downward by stages through the oxidized zone and through the supergene sulphide zone so that a considerable concentration may eventually be reached.

Chalcopyrite is readily attacked by oxygen and by ferric sulphate. It is slightly attacked by dilute sulphuric acid.

$$CuFeS_2+8O=CuSO_4+FeSO_4.$$

As the ferrous salt is easily transformed into limonite, pseudomorphs of that mineral after chalcopyrite are extremely common.

Bornite is more strongly attacked by dilute sulphuric acid than chalcopyrite,[1] but like all copper sulphides easily decomposed by ferric sulphate.

$$Cu_5FeS_4+2H_2SO_4+18O=5CuSO_4+FeSO_4+2H_2O.$$

Chalcocite like covellite is very slightly attacked by dilute

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[1] E. G. Zies, E. T. Allen and H. E. Merwin, Some reactions involved in secondary copper sulphide enrichment, *Econ. Geol.*, vol. 11, 1916, p. 476.

sulphuric acid but is decomposed by ferric solutions which transform it to sulphates, probably also to covellite:

$$Fe_2(SO_4)_3 + Cu_2S = CuSO_4 + 2FeSO_4 + CuS.$$

$$Fe_2(SO_4)_3 + CuS + 3O + H_2O = CuSO_4 + 2FeSO_4 + H_2SO_4.$$

$$Cu_2S + FeCl_3 = FeCl_2 + CuCl + CuS.$$

Enargite and tetrahedrite are likewise slowly decomposed by dilute solutions of ferric sulphate. Normally the antimony remains as insoluble oxide while the arsenic is carried away unless fixed as arsenates of copper by carbonate solutions or limestone. The universally resulting cupric sulphate[1] is more or less completely fixed as malachite, azurite and brochantite, chryscolla and similar products by solutions containing carbonates or silica. Ultimately even these minerals will be leached for they are slightly soluble in water containing carbon dioxide and easily soluble in dilute sulphuric acid. They may, however, be reduced to cuprite and the cuprite to native copper which again may go into solution with H_2SO_4.

The oxidation products of the chalcocite zone are described on p. 858.

Supergene Copper Sulphides.—So far we have considered the products of direct oxidation, in the uppermost part of pyrite deposits; they form under the influence of acid solutions containing free oxygen and ferric salts; limonite and oxy-salts of copper result, mixed with residual quartz. At a certain depth, usually at the water level, if that remains in its original horizon, the material changes from a brown to grey or bluish color and copper sulphides begin to appear. At first they form pulverulent or sooty masses with little residual pyrite. Their quantity gradually decreases; we find grains of pyrite, chalcopyrite, bornite, or zinc blende covered by coatings of covellite or chalcocite and the microscope gives evidence that the process is a replacement of primary sulphides by the two minerals mentioned (Figs. 274, 275, 276). Occasionally secondary chalcopyrite or bornite will appear but no secondary pyrite. The alteration proceeds from reticulating veinlets or in concentric shells. Below the upper part of the zone the chalcocite may be compact with dark

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[1] Cuprous sulphate may form as intermediate product in some reactions but it is unstable and according to W. H. Emmons has not been discovered in any analysis of mine waters.

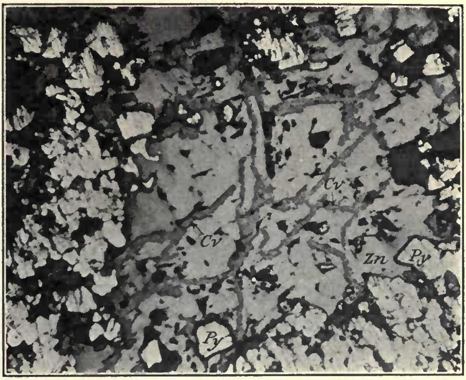

FIG. 274.—Pyrite (*Py*) intergrown with zinc blende (*Zn*), which is altering to covellite (*Cv*), Kyschtim, Russia. Magnified 45 diameters. *After L. C. Graton and J. Murdoch.*

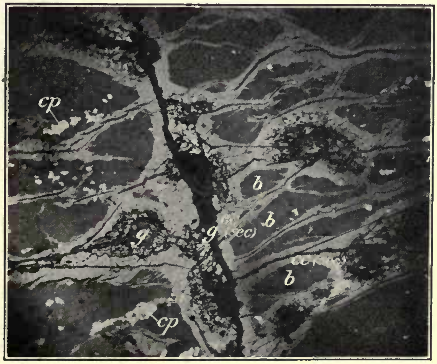

FIG. 275.—Bornite (*b*) altering along fissures to chalcopyrite (*cp*), Ajo, Arizona. *g*, Gangue; *cc* (*sec*), secondary chalcocite. Magnified 50 diameters. *After L. C. Graton and J. Murdoch.*

grey metallic luster. Veins of pyrite may be converted to sooty or massive chalcocite with only few residual grains of the original mineral. In some deposits large and rich masses of chalcocite, more rarely covellite, may be formed in this manner.

Kaolin may form along with the secondary sulphides partly at the expense of the sericite; other minerals are chalcedonic and opaline silica, also alunite. Pyrite is unstable in this zone. Deposition of quartz is very unusual.

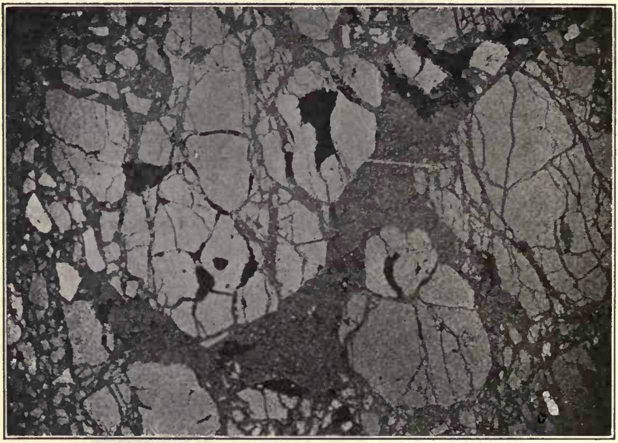

FIG. 276.—Pyrite altering to chalcocite (dark); black is open field; Clifton, Arizona. Magnified 30 diameters.

The upper limit of the chalcocite zone is usually sharp. In depth, the secondary sulphides may cease equally suddenly (Fig. 277), but it is more common to find a gradual decrease. Chalcocite zones of wide extent are usually explored by churn drilling and the plotting by graphic methods of the assays obtained gives a clear idea of the sharp changes or gradual transitions due to the enrichment.[1] The depth of chalcocite zone often reaches 1,000 feet and in exceptional cases considerably more. On the

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[1] E. H. Perry and A. Locke, Interpretation of assay curves for drill holes, *Trans.*, Am. Inst. Min. Eng., vol. 54, 1917, pp. 93–99.

other hand, the enrichment may be confined to a thin layer of a thickness of a few feet only (Fig. 278).

The development of the chalcocite zones is dependent on climate and water level as well as on composition and texture of the primary rocks and ores. Permeable rocks like sericitized granite, porphyry or schists are particularly favorable environments. In compact contact-metamorphosed shales and limestones, the secondary sulphides do not readily develop. In limestone the zone is usually irregular and shallow because

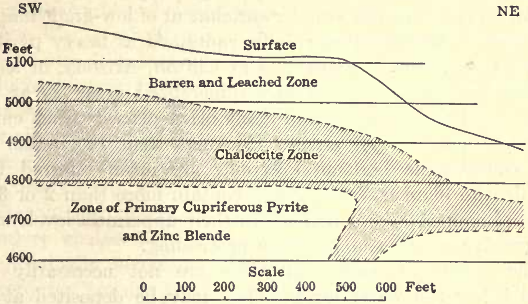

Fig. 277.—Longitudinal section of chalcocite zone in a vein at Morenci, Arizona.

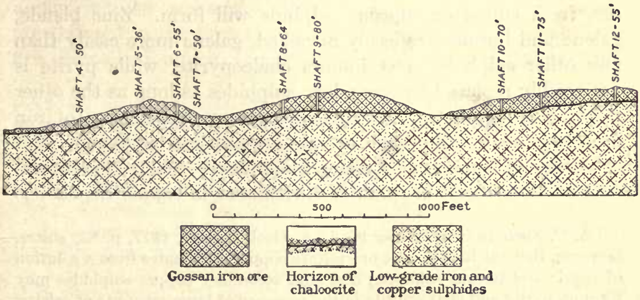

Fig. 278.—Chalcocite zone at Ducktown, Tennessee. *After W. H. Emmons, U. S. Geol. Survey.*

the basic copper carbonates here form so easily.[1] Wherever the chalcocite zone is present a marked enrichment has taken place.

Chalcocite zones may develop in primary ores of economic value. They may also form by enrichment of low-grade material (protore, p. 833) whether this be contained as heavy pyrite in veins of low tenor in copper, as at Clifton, Arizona; or as disseminations of pyrite in larger mineralized areas as at Ely, Nevada, Miami, Arizona, and many other places. Such enrichments, usually of not greater thickness than 100 to 300 feet, but considerable horizontal extent, are often referred to as "chalcocite blankets" and rarely contain more than 2 or 3 per cent. of copper, except perhaps in their uppermost levels where progressive enrichment has been proceeding.

The secondary copper sulphides are not necessarily confined below the water level. They may be deposited at any place in the oxidized zone where there is a deficiency in oxygen and ferric sulphate, as well shown at Tintic, Utah, and other places.[2] Such supergene sulphides are, however, likely to be spotted and irregular in occurrence.

A lowering of the water level and resulting oxidation of the chalcocite causes a progressive enrichment and an enlargement of the zone for wherever the cupric sulphate reaches the primary ore, fresh cupric or cuprous sulphide will form. Zinc blende, galena and bornite are easily attacked, galena more easily than any other sulphide; next follows chalcopyrite, while pyrite is not readily replaced by secondary sulphides as long as the other minerals are present. The whole process involves removal of iron on a large scale. Zinc and arsenic are also carried away.

Theory of Supergene Copper Sulphides.[3]—The recent literature on the subject of sulphide enrichment in copper deposits is

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[1] A. C. Spencer (*Prof. Paper* 96, U. S. Geol. Survey, 1917, p. 82) states, however, that calcite does not precipitate copper carbonates from a solution of cupric and ferrous sulphate; also that secondary copper sulphides may form on pyrite and chalcopyrite in the presence of large amounts of calcite.

[2] W. Lindgren, *Econ. Geol.*, vol. 10, 1915, p. 236.

[3] H. N. Stokes, On the solution, transportation and deposition of copper, silver and gold, *Econ. Geol.*, vol. 1, 1906, pp. 644–650. Also *Idem*, vol. 2, 1907, pp. 12–23.

E. Posnjak, E. T. Allen and H. E. Merwin, The sulphide ores of copper, *Econ. Geol.*, vol. 10, 1915, pp. 491–535.

E. G. Zies, E. T. Allen and H. E. Merwin, Some reactions involved in voluminous and it is not possible to follow it in detail in this place. The deposition is governed by Schürmann's reactions (p. 843) so that in general the simple sulphides of copper replace the simple sulphides of iron, lead and zinc. They also replace the copper-iron sulphides like chalcopyrite and bornite and the sulphantimonides like tetrahedrite and the sulpharsenides like tennantite and enargite. There is not much evidence of deposition of copper sulphides by precipitation by hydrogen sulphide or by alkaline sulphides though no doubt such reactions may also take place, especially when pyrrhotite is one of the primary minerals. Spencer has pointed out that the action of cupric sulphate on sulphides is really a process of oxidation of iron, and he as well as Graton and others have suggested that the formation of secondary chalcocite probably involves a series of transitions with gradually increasing richness of copper, such as pyrite, chalcopyrite, bornite, covellite and chalcocite. All the stages are rarely observed together and the last two minerals are surely the most important. The extent of bornite as a secondary mineral is as yet in doubt.

Near the surface the mine waters are solutions of sulphuric acid and ferric sulphate; in depth their acidity decreases and ferrous sulphate increases; at greater depth the waters become neutral and finally alkaline.[1] Cupric sulphate is present all along but the secondary sulphides are evidenlty not readily precipitated in the presence of much ferric sulphate though they are normally found in the presence of ferrous sulphate of slight

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secondary copper sulphide enrichment. Contribution No. 7. Secondary enrichment investigation, *Econ. Geol.*, vol. 11, 1916, pp. 407–503.

C. F. Tolman, Jr., Secondary sulphide enrichment, *Min. & Sci. Press*, vol. 106, 1913, pp. 38–43, 141–145, 178–181.

C. F. Tolman, Jr., Observations on certain types of chalcocite, etc., *Trans.*, Am. Inst. Min. Eng., vol. 54, 1917, pp. 402–442.

L. C. Graton and J. Murdoch, The sulphide ores of copper, *Trans.*, Am. Inst. Min. Eng., vol. 45, 1914, pp. 126–181.

A. C. Spencer, Geology and ore deposits of Ely, Nevada, *Prof. Paper* 96, U. S. Geol. Survey, 1917, pp. 76–91. Excellent review of subject.

W. H. Emmons, The enrichment of ore deposits, *Bull.* 625, U. S. Geol. Survey, 1917, pp. 154–249. Excellent and complete review. Bibliography on pp. 20–33.

[1] G. S. Nishihara, The rate of reduction of acidity by descending waters, etc., *Econ. Geol.*, vol. 9, 1914, p. 743–757.

F. F. Grout, On the behavior of cold, acid solutions, etc., *Econ. Geol.*, vol. 8, 1913, p. 429.

acidity. The early work by H. V. Winchell, C. F. Tolman, Jr., H. N. Stokes, E. C. Sullivan, T. T. Read, T. W. B. Welsh, C. A. Stewart and A. C. Spencer showed that secondary sulphides of copper may replace pyrite and other sulphides in cupric sulphate solution. In 1906 Stokes had established quantitatively the reaction with pyrite. In the notable paper by E. G. Zies, E. T. Allen and H. E. Merwin of the Geophysical Laboratory all the various reactions were quantitatively determined at temperatures of 40° C. and 200° C.

The reaction with sphalerite is as follows:

$$ZnS + CuSO_4 = CuS + ZnSO_4.$$

The presence of sulphuric acid accelerates the reaction. When cupric sulphate and galena react at 35° C. cupric sulphide is first formed which is further attacked by cupric sulphate yielding cuprous sulphide. The attack on chalcopyrite at 40° and at 200° C. is expressed by the equation

$$CuFeS_2 + CuSO_4 = 2CuS + FeSO_4.$$

In this reaction the cupric sulphide again alters to cuprous sulphide on further attack by $CuSO_4$. The presence of sulphuric acid does not retard the reaction.

The action between bornite and cupric sulphate at the same temperature is expressed by the equations

$$5Cu_5FeS_4 + 11CuSO_4 + 8H_2O = 18Cu_2S + 5FeSO_4 + 8H_2SO_4.$$

$$Cu_5FeS_4 + CuSO_4 = 2Cu_2S + 2CuS + FeSO_4.$$

Bornite is attacked by H_2SO_4 resulting in CuS and Cu_2S, and $FeSO_4$, hydrogen sulphide developing at the same time. These products will react and form secondary chalcopyrite.

Pyrite alters to chalcocite and covellite according to Stokes' formula:[1]

$$5FeS_2 + 14CuSO_4 + 12H_2O = 7Cu_2S + 5FeSO_4 + 12H_2SO_4$$

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[1] Stokes verified this reaction at 80° and 100° C. with neutral solution. Some CuS was also formed, less at 100° than at 180°. Cuprous sulphate also forms as an intermediate product. Cupric and ferrous sulphate mix in all proportions without change, except at high temperatures (200° C.), when, according to Stokes, cuprous sulphate and ferric sulphate form; the latter hydrolyzes to ferric hydrate and H_2SO_4, while cuprous sulphate deposits copper upon cooling. This reaction is not likely to take place at temperatures ordinarily existing in the oxidized zone.

The formation of covellite is expressed by the following formula:

$$4FeS_2+7CuSO_4+4H_2O=7CuS+4FeSO_4+4H_2SO_4.$$

Sulphuric acid exerts a markedly retarding influence on these reactions.

According to Zies, Allen and Merwin pyrrhotite alters to chalcopyrite and probably later to bornite when attacked by cupric sulphate but the reaction was not followed quantitatively. Most observers have assumed that covellite is earlier than chalcocite and Zies, Allen and Merwin confirm this experimentally:

$$5CuS+3CuSO_4+4H_2O=4Cu_2S+4H_2SO_4.$$

It is probable that this reaction is reversible for in many cases covellite is an alteration product of chalcocite.

According to Posnjak, Allen and Merwin[1] cupric sulphide is formed when cuprous sulphide is treated with dilute acid solution in the presence of oxygen. This probably explains the development of covellite in oxidizing chalcocite and its occasional crystallization together with products of oxidation like anglesite.

The Relation of Chalcocite, Covellite and Bornite.—According to the authors just cited all chalcocite crystals so far found—they are not common—crystallize in the rhombic system, while all synthetic chalcocite formed above 91° C. is isometric. They also found that chalcocite may hold covellite in solid solution and that above 8 per cent. covellite there is no inversion point. Chalcocite when pure is white in reflected light but many varieties are more or less bluish, and this is held to be caused by dissolved CuS. While this is undoubtedly correct, it is true that some of the "blue chalcocites" under high objectives are resolved into a mixture of chalcocite with covellite. Some even show pinkish mottling and are found to contain particles of bornite and chalcopyrite. The etch figures of normal low temperature chalcocite often show a network of three partings, suggesting isometric symmetry. This has been investigated by L. C. Graton and J. Murdoch, and C. F. Tolman, Jr., who suggest that this parting may have been inherited from the bornite from which the chalcocite was derived. The question of "primary chalcocite" has occupied much space in the literature of late. In so far as the view of chalcocite as a hypogene mineral in ore

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[1] *Op. cit.*, p. 528.

deposits is based on the regular "intergrowth" of chalcocite and bornite it is decidedly untenable for these "intergrowths" are surely replacements by chalcocite and could well have been formed by supergene solutions. The problem is not solved but it may be said that chalcocite has not yet been proved a primary or hypogene mineral in sulphide deposits of hypogene origin.[1]

L. C. Graton,[2] J. C. Ray[3] and others have also held that covellite may be of hypogene origin because it replaces, in crystal form, other sulphides like pyrite, enargite and sphalerite. However, covellite formed by undoubtedly supergene solutions always shows an exceedingly strong tendency to develop blades and crystals. Covellite has not yet been proved a primary or hypogene mineral in sulphide deposits of hypogene origin.

Oxidation of Chalcocite Zones.—Where erosion and water level have been stationary for a long time no changes take place in the chalcocite zone except by the gradual increase due to continued slow leaching of the oxidized zone. But when erosion is quickened or the water level subsides the reactions of oxidation may invade the chalcocite zone. The gossan may be entirely removed and then oxidation will be working on enriched sulphide ore in which there will be comparatively little iron as pyrite. Other elements like arsenic may also have been removed during chalcocitization. The oxidation of such materials may result in almost complete leaching of copper and a residual outcrop consisting of quartz with some sericite in which even copper stains may be lacking. Such conditions exist at Clifton, Miami, and many other places in Arizona. The outcrops at Butte represent also a leached chalcocite zone and contain little copper.

The course of the oxidation depends largely on the amount of residual pyrite. Where much of this is present the chalcocite is decomposed to cupric sulphate while pyrite protected by the chalcocite remains a little longer until finally decomposed into

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[1] For discussion regarding this subject see B. F. Laney, *Econ. Geol.*, vol. 6, 1911, pp. 399–411 (Virgilina district).

A. F. Rogers, *Econ. Geol.*, vol. 8, 1913, pp. 781–795 (Butte).

H. W. Turner and A. F. Rogers, *Econ. Geol.*, vol. 9, 1914, pp. 359–391 (Engel's mine, Cal.).

L. C. Graton and D. H. McLaughlin, *Econ. Geol.*, vol. 12, 1917, pp. 1–38 (Engel's mine).

J. F. Tolman, Jr., *Econ. Geol.*, vol. 12, 1917, pp. 379–386.

[2] *Trans.*, Am. Inst. Min. Eng., vol. 45, 1914, p. 51.

[3] *Econ. Geol.*, vol. 9, 1914, p. 473.

iron sulphate. When there is little or no pyrite the chalcocite normally changes by oxidation to covellite and cuprite, the small amount of sulphuric acid and ferric sulphate available seems to be sufficient to dissolve these (Fig. 279). Again, the chalcocite may change to brochantite which seems to be particularly common in oxidizing supergene sulphide zones, or more rarely to malachite and chrysocolla.

$$2Cu_2S+O=2CuS+Cu_2O.$$

$$2Cu_2S+10O+4H_2O=H_6Cu_4SO_{10}\text{ (brochantite) }+H_2SO_4.$$

The cuprite is often reduced by ferrous sulphate to native copper and this may again be dissolved by sulphuric acid. At times chalcocite is directly reduced to native metal which may preserve the structure of the black sulphide.[1] This might have been effected by ferric sulphate according to the following reaction, and probably in many other ways.

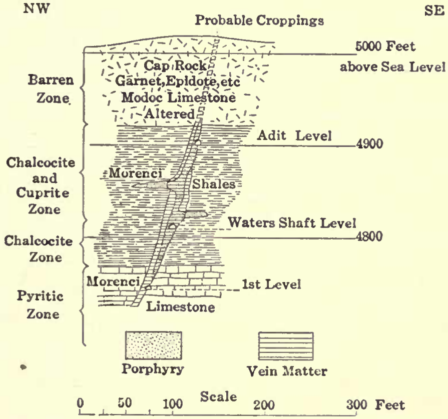

FIG. 279.—Secondary zones in copper veins in contact-metamorphic rocks Clifton, Arizona.

$$Cu_2S+3Fe_2(SO_4)_3+4H_2O=2Cu+6FeSO_4+4H_2SO_4.$$

In the chalcocite blankets it is not common to find much native

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[1] W. Lindgren, *Prof. Paper* 43, U. S. Geol. Survey, 1905, p. 101.

metal, but at Chino, New Mexico, most of the secondary sulphide appears to have been converted to copper.[1]

EXAMPLES OF OXIDATION OF COPPER DEPOSITS

General Features.—The study of the various modes of enrichment in copper deposits is a subject full of difficulties. We find the most diverse development even in a region of uniform general climate. Take, for instance, the Sonora-Arizona province, where the rainfall is small and the climate warm. At Los Pilares, Sonora, near Nacozari, a gossan of barren hematite 100 feet deep is underlain by an ill-defined zone with bornite and chalcocite, changing below the 500-foot level to primary chalcopyrite-pyrite ore. In other parts of Sonora, according to Finlayson,[2] are gossan and chrysocolla ores extending to a depth of 200 to 400 feet; below this is a shallow zone of secondary sulphides. Again, at Clifton, Arizona, there are in the contact-metamorphic deposits in limestone strong gossans, sometimes rich in copper, underneath which no secondary sulphides are found. Pyritic veins in porphyry at the same place have a barren siliceous outcrop without gossan and perhaps 150 feet thick, below which lies a rich chalcocite zone that in a few hundred feet or less changes to lean primary sulphides. Other veins near by show chrysocolla from the surface down to a shallow chalcocite zone at 100 feet. At Miami, Arizona, where enrichment has taken place through concentration in large masses of pyritized and sericitized rocks, there is a thick, almost barren zone of oxidation below which, at depth of 200 to 1,100 feet, lies a blanket of chalcocitized rock from 50 feet or less to 300 feet in thickness. Almost everywhere in Arizona the chalcocite zone is far above the water level, though according to the accepted theory the chalcocite zone should form at and below the water level. The water-table may formerly have been higher and coincided with the top of the chalcocite zone, but this cannot always be proved.

In the normal course of oxidation a gossan must form and the three zones should be distinct; if the gossan is not present it has been eroded and the barren upper zone has then been formed by leaching of the zone of sulphide enrichment, the copper solution

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[1] L. C. Graton, *Prof. Paper* 68, U. S. Geol. Survey, 1910, p. 316.
Sidney Paige, *Econ. Geol.*, vol. 7, 1912, pp. 547–559.

[2] A. M. Finlayson, Economics of secondary enrichment, *Min. and Sci. Press*, July 16 and 23, 1910.

descending to further enrich the deposit in depth. In regions of deep erosion it is exceedingly rare to find a strong chalcocite enrichment in deposits exposed in the lower parts of the canyons. In glaciated or rapidly eroded regions almost all enrichments may be lacking.

Rio Tinto.—The pyritic deposits of Rio Tinto, southern Spain,[1] are situated in a country of sub-tropical climate, with an annual rainfall of about 25 inches and mature topography, where erosion makes slow headway. The primary deposits are thick lenses of pyrite containing less than 1 per cent. of copper. There is a heavy gossan of massive hematite, 45 to 90 feet thick, containing no copper, over 50 per cent. iron, and 10 to 15 per cent. of siliceous and argillaceous matter. The depth of oxidation has everywhere been determined by the ground-water level. The lower limit of the gossan is sharp and the line is often marked by a thin earthy zone with notable quantities of gold and silver,[2] believed to represent an enrichment caused by leaching of the gossan by solutions containing chlorine and ferric sulphate. The top of the sulphide zone for a thickness of a few feet is composed of leached pyrite with a trace of copper, resembling the upper part of the chalcocite zone of Morenci, Arizona. Below this begins the zone of enriched sulphides, containing in the upper part 3 to 12 per cent. copper, gradually becoming poorer downward and passing into lean pyritic ore assaying 1 per cent. or less of copper. The depth at which the unaltered ore is reached ranges from 200 to 1,500 feet below the outcrop. The enriched pyrite contains mainly chalcocite, but also some secondary chalcopyrite. The bulk of the Rio Tinto copper production to-day is derived from enriched ore.

Mount Morgan.[3]—The great gold and copper deposit of Mount Morgan, in Queensland, which since 1886 has yielded about $65,000,000 in gold and now bids fair to become a great copper-producing property, is most interesting and shows the peculiar feature of great gold enrichment with almost entire absence of a

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[1] A. M. Finlayson, The pyritic deposits of Huelva, Spain, *Econ. Geol.*, vol. 5, 1910, pp. 357–372; 403–437.

[2] J. H. L. Vogt, *Zeitschr. prakt. Geol.*, 1899, p. 250.

[3] J. M. Maclaren, Gold, London, 1908, pp. 333–337.

W. F. Gaby, Petrography of the Mt. Morgan mine, *Trans.*, Am. Inst. Min. Eng., vol. 55, 1917, pp. 263–283.

J. F. Newman and J. F. C. Brown, *Trans.*, Austral. Inst. Min. Eng., vol. 15, pt. 2, 1910.

zone of secondary copper ores. The region has a tropical climate and moderate rainfall; the topography is of the moderately mature type. The water level is probably deep. The irregular deposit is apparently a replacement in Carboniferous rocks, surrounded on both sides by intrusive granite, thus recalling the deposits worked by the Reforma mine, in Guerrero, Mexico, and the Mount Lyell mine, in Tasmania.

At the outcrop there was an extremely rich zone with free gold in kaolin, limonite, and black manganese. Below this was found a zone of a cellular, almost pumiceous siliceous mass, evidently a quartz skeleton resulting from the removal of pyrite; this was poorer in gold, but the kaolin that was in places associated with it was rich in silver. The sharply defined lower limit of the oxidized ore was met at 180 to 300 feet below the surface, and the primary ore consisted at first of pyrite, then of pyrite with chalcopyrite, carrying 2 to 3 per cent. copper and $1 to $8 in gold to the ton. It is difficult to account for the lack of a chalcocite zone. Unquestionably there has been concentration of gold on a large scale at the surface, probably caused by the presence of unusual amounts of chlorine. It is noteworthy that the gold has been precipitated mainly at the surface and could not be carried down into lower levels.

Butte.—The copper deposits of Butte, Montana,[1] form a system of east-west, steeply dipping veins cutting quartz monzonite as well as dikes of aplite and granite porphyry (Modoc porphyry) intrusive into this granular rock. They are mainly disseminated pyritic replacements along fissures and contain mainly pyrite with enargite, bornite, chalcocite, zinc blende and a little chalcopyrite and covellite. There is a scant gangue of quartz.

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[1] S. F. Emmons, W. H. Weed, and G. W. Tower, Jr., Butte folio (No. 38), *Geol. Atlas* U. S. Geol. Survey. W. H. Weed, Geology and ore deposits of the Butte district, Mont., *Prof. Paper* 74, U. S. Geol. Survey, 1912. J. F. Simpson, *Econ. Geol.*, vol. 3, 1909, pp. 628–636. See also Reno H. Sales, Superficial alteration of Butte veins, *Econ. Geol.*, vol. 5, 1910, pp. 15–21; vol. 3, 1908, pp. 326–331. Ore deposits of Butte, Montana, *Trans.*, Am. Inst. Min. Eng., vol. 40, 1914, pp. 3–106. C. T. Kirk, Conditions of mineralization in the copper veins at Butte, *Econ. Geol.*, vol. 7, 1911, pp. 35–82. J. C. Ray, Paragenesis of the ore minerals in the Butte district, *Econ. Geol.*, vol. 9, 1914, pp. 463–481. D. C. Bard and H. M. Gidel, Mineral associations in the Butte district, *Trans.*, Am. Inst. Min. Eng., vol. 46, 1914, pp. 123–127. A. P. Thompson, The occurrence of covellite at Butte, *idem*, vol. 52, 1916, pp. 563–596. W. W. Atwood, The physiographic conditions at Butte, etc., *Econ. Geol.*, vol. 11, 1916, pp. 697–740.

Extensive sericitization along the veins is characteristic. These veins are cut by a system of veins trending northwest and characterized by much enargite besides the other minerals mentioned. Finally there are northeastward-trending veins that have caused dislocations of the older veins amounting in places to several hundred feet. The supergene ores are found mainly in the first two classes of fractures and consist of chalcocite with smaller amounts of covellite. Many of the fault veins carry chalcocite as drag but owing to their clayey character the circulation along them has been sluggish and they do not contain large masses of ore (Figs. 46 and 47).

The outcrops are not prominent and the copper is leached from them; in some places they contain chrysocolla. The depth of this oxidized barren zone of honeycombed quartz veins ranges from 10 to 400 feet. In the central copper area, where the granite is greatly altered, the upper limit of the sulphides is practically a plane in spite of surface inequalities of nearly 300 feet, evidently depending more upon the alteration of the country rock than on the topography. In the leached zone there is a slight enrichment of silver, the material containing as much as 30 ounces per ton, in contrast to 2 5 ounces per ton in the sulphide ore. No secondary silver sulphides have been noted. The gold tenor is uniform throughout the copper area, indicating that practically no secondary concentration has taken place. The ores average 30 to 50 cents per ton in gold, with little difference between oxidized ore and that of the sulphide zone. A sharp line of demarcation separates the two zones, the change in many places occurring within 2 or 3 feet vertically. At this level the ores contained much chalcocite and averaged 8 per cent. or more of copper. Solid masses of glance ore, 15 feet or more in thickness, were found. Covellite is less abundant and secondary chalcopyrite is rare. Corroded crystals of chalcocite and others coated by quartz are mentioned by H. V. Winchell. In depth the enriched ore gradually decreases in value, but low-grade ore of about 2.5 per cent. persists to depths of 3,000 feet or more, particularly along planes where the circulation of water was energetic. In general, chalcocitization in the upper levels was accompanied by kaolinization.

It is generally conceded that the upper exceptionally rich parts of the veins contained an abundance of supergene chalcocite (Fig. 47). Sales designates these as the "sooty" chalcocite zone implying its soft and pulverulent character. However, much of that ore mined in the early days was compact and showed bornite.

Below this, enargite began to come in but the chalcocite still persisted and is even now found in the deepest levels (3,500 feet below the surface) though there is much pyrite and enargite and some tetrahedrite. Covellite occurred in considerable amounts in the upper levels and as deep as 2,000 feet or more. Many observers, among them Reno Sales, J. C. Ray and A. P. Thompson believe that this deeper chalcocite is mainly of hypogene origin though later than the first succession, which is represented by quartz, pyrite and enargite. The bornite is generally older than the chalcocite which even in depth seems to be the later mineral. Almost any specimen of the older, rich chalcocite ore shows ill-defined remnants of bornite. The observers mentioned above also believe that the covellite may be of hypogene origin.

There is no doubt that the Butte ores show repeated replacements of a complicated order, but I am inclined to believe that the hypogene origin of chalcocite, covellite, and perhaps bornite has not been proved. As far as can be judged the chalcocite is of the rhombic modification which is only stable below 91° C., a low temperature which can hardly have been reached during hypogene condition in the Butte veins. It must be remembered that wherever cupric sulphate is carried supergene sulphides may form and there is evidence of a distinct downward movement of the groundwater far below the water level.

I am inclined to doubt the hypogene origin of the bornite remnants in the chalcocite. Even those who, like Ray, believe in the hypogene origin admit that it is a secondary mineral. I have seen exactly the same association in the small veins of Chuquicamata (for instance at the Emilia mine) which in their lower levels carry only pyrite-enargite ore.

According to this view the Butte veins contain a hypogene pyrite-enargite ore and in them has been developed an exceedingly strong chalcocite zone with elimination of arsenic in the upper levels. The oxidized zone as well contains no arsenic and has been formed by oxidation of the chalcocite zone which, below, became progressively enriched.

In explanation of the deep chalcocite zone Emmons and Weed state that the block in which the veins are contained has been faulted down probably several thousand feet and that thus the water level might formerly have stood lower than at the present time. This explanation is not satisfactory, for it is not probable that the water level in this region has ever been at a considerable depth below the surface, and in the adjacent and higher block on the east the water level still remains high. It is conceded that the present barren zone was created by leaching of the former upper part of the chalcocite zone and that during this process the copper was sharply concentrated in the upper part of the present enriched zone. Therefore the latter, like many other bodies of secondary chalcocite, must be of considerable geological age and long antedate the faulting. When it was accumulated the water level was assuredly much higher than at present. This view is supported in a notable paper by Atwood in which physiographic principles are applied to the study of supergene enrichment.

Ely.—An important deposit of secondary chalcocite is now being worked on a large scale at Ely, Nevada,[1] by the Nevada Consolidated Copper Company.

The reserves of the company are estimated at 50,000,000 tons, containing from 1.5 to 2 per cent. of copper. The ore carries also, in ounces per ton, 0.018 in gold and 0.088 in silver.

The geological relations are similar to those of the Arizona deposits. Intrusions of monzonite porphyry in Paleozoic limestone caused contact metamorphism of the limestone, silicification of both rocks, and some development of copper deposits, few of which are of economic importance. After intrusion the porphyry became impregnated with disseminated pyrite with a little chalcopyrite, the silicates being altered to sericite and pyrite. When the intrusive masses became exposed by erosion to the action of oxidizing waters a downward migration of copper sulphate, either from the porphyry itself or from the overlying contact deposits, effected a chalcocitization over wide areas.

The leached zone is from 50 to 200 feet in depth and forms an iron-stained soft mass, in places containing oxidized copper ores; below this lies the chalcocite zone, consisting of white earthy porphyry with disseminated grains and flakes of chalcocite and a little pyrite. This zone has a maximum depth of about 500 feet, the copper minerals gradually diminishing down-

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[1] A. C. Spencer, *Prof. Paper* 96, U. S. Geol. Survey, 1917.

ward to the pyritic valueless protore; the upper limit of the chalcocite is rather sharply defined. Water is beginning to come in a depth of 100 feet, though the general water level in the porphyry is said to be 385 feet below the surface (Fig. 280).

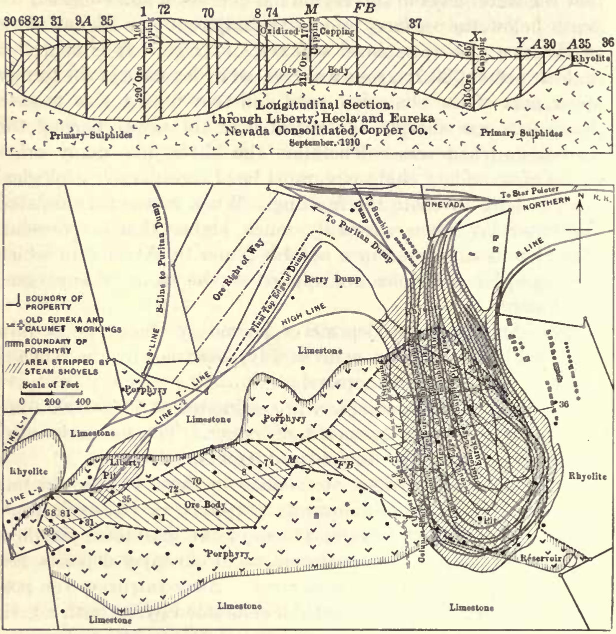

FIG. 280.—Plan and section of chalcocite deposit at Ely, Nevada. From annual report of Nevada Consolidated Copper Co.

Bingham.—Relations similar to those at Ely exist at Bingham, Utah,[1] in a region of much sharper relief and medium aridity. A small mass of monzonite is here intruded into Carboniferous (Pennsylvanian) quartzite and limestone, and has apparently caused the rich mineralization of the Bingham deposits. A large part of the monzonite, one mile in length and half a mile

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[1] J. M. Boutwell, *Prof. Paper* 38, U. S. Geol. Survey, 1905.

J. J. Beeson, The disseminated copper ores, Bingham Canyon, Utah, *Trans.*, Am. Inst. Min. Eng., vol. 54, 1917, pp. 356–401.

in width has been subjected to hydrothermal metamorphism resulting in the development of disseminated pyrite, chalcopyrite and bornite. This "protore" which probably contains 1 per cent. copper or less has been enriched by supergene solution, depositing chalcocite and, in its upper part, also covellite, resulting in a low-grade ore containing 1.5 per cent. copper and, in ounces per ton, 0.018 (37 cents) in gold and 0.25 (25 cents) in silver. The output of ore per day is almost 30,000 tons; steam shovels are now used almost exclusively. Below a leached surface zone (with some oxidized ore) about 70 feet in depth, lies the chalcocite blanket. The total thickness of the enriched zone is not fully determined. The developed ore amounts to about 360,000,000 tons (Fig. 281).

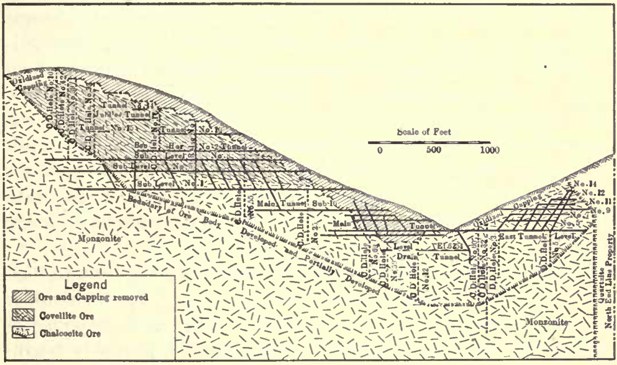

FIG. 281.—Longitudinal section through central portion of ore-body of the Utah Copper Co., Bingham Canyon, Utah.

Beeson has shown that some supergene chalcopyrite may be formed as an intermediate product between pyrite and chalcocite, and bornite as an intermediate mineral between chalcopyrite and chalcocite. Bornite and chalcopyrite may also be developed in reverse order by replacement of chalcocite and covellite.

The Southwestern Chalcocite Deposits.—In the arid country of southern Arizona and New Mexico we find an interesting group of secondary sulphide deposits similar to the last two examples given. They are sometimes called chalcocite blankets or disseminated chalcocite deposits, and excellent representatives of them are found at Clifton, Globe, Ray, and Santa Rita, and in the Burro Mountains. In brief, the concentration has been proceeding in porphyry, granite, or schist containing disseminated pyrite with a little chalcopyrite. Enrichment through replacement of pyrite by chalcocite has in places occurred along fissures or fissured zones, or still more commonly in irregular areas of fractured and brecciated rocks. The result is a chalcocite ore containing 2 to 4 per cent. copper and also some residual pyrite; this zone is from 100 feet or less up to several hundred feet in thickness. Above it lies a barren oxidized and leached zone reaching to the surface and from 50 to 1,000 feet in thickness; in places this zone contains some oxidized ore. Below the chalcocite, the primary pyritic dissemination extends to an unknown depth, the rock containing but a fraction of a per cent. of copper. The upper limit of the chalcocite zone is sharply defined; the richest ore is found here, gradually decreasing in tenor as depth increases. The water level usually lies at or below the lower limit of the chalcocite zone, and the zone itself, or at any rate the top of it, is for the most part high above the present drainage level.

Evidently the secondary sulphides could not have been formed in their present places under present conditions, for their upper parts are now being actively oxidized. They give evidence of having been accumulated during a long period, probably beginning in the late Tertiary, when the climate was damp and the water level high, before erosion had cut to its present depth. The overlying lean porphyry was leached of its scant copper content, the copper descending as sulphate to become precipitated as chalcocite on the primary pyrite in depth. At some places the copper solutions may have been derived partly from once overlying, now eroded contact-metamorphic deposits.

These deposits are then old—marooned, as it were, high above their normal position and in an unstable condition. Probably they were once thicker and poorer than now and covered by a gossan. Erosion has carried away the surface gossan, and the scant rain waters have leached the upper part of the underlying chalcocite zone—now the barren zone—and driven the copper downward to replace the remaining pyrite at the level in the zone where the oxygen of the descending water became exhausted. Thus is explained the richness near the top, and it follows as a corollary that chalcocite may be deposited above the permanent water level, provided not much oxygen is present.

Globe.—The copper deposits at Globe, Arizona, and the geology of the surrounding region have been described by Ransome.[1] A few miles from Globe, in a region of moderate relief, there is an area of granite (Schultze granite) intrusive into the pre-Cambrian Pinal schist; in the latter near the contact several disseminated chalcocite deposits have lately been discovered.

At the Miami mine[2] the leached zone is about 200 feet deep and contains in places oxidized ores; a sharp line of demarcation separates it from the underlying chalcocite. The deposit forms a flattened mass which in depth gradually increases in extent. On the 270-foot level the chalcocite area occupies 1 acre; on the 370-foot level, 3 acres; on the 470-foot level, 16 acres. The average tenor of the ore is over 3 per cent. of copper near the top of the chalcocite zone, but falls to 2.65 per cent. on the 570-foot level. At greater depth the percentage of copper in the ore changes abruptly from 2 per cent. to 1 per cent. or less. The mine produces a little water on the 450-foot level.

At the neighboring Inspiration mine the leached surface zone is from 50 to 575 feet deep, averaging 367 feet, while the underlying enriched zone averages 155 feet in thickness.[3] It is stated that 21,000,000 tons of ore containing 2 per cent. copper has been developed over an area of 40 acres. To develop this ore 81 holes were drilled and underground development work aggregating 27,500 feet was done. In the whole ore zone, it is said, 80,000,000 tons of 2 to 2.5 per cent. ore has been developed.

These deposits are thought to have been formed during the last part of the Tertiary period. Their oxidation is now in progress, with enrichment and concentration of the underlying chalcocite.

Ray.—At Ray Arizona,[4] about 25 miles southwest of Globe, a similar but more extensive chalcocite blanket has been dis-

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[1] F. L. Ransome, *Prof. Paper* 12, U. S. Geol. Survey, 1903.

[2] C. F. Tolman, *Min. and Sci. Press*, Nov. 13, 1909; E. Higgins, *Eng. and Min. Jour.*, April 9, 1910; R. L. Herrick, *Mines and Minerals*, July, 1910; F. L. Ramsome, *Prof. Paper* (in press), U. S. Geol. Survey.

[3] Recent developments have shown that the chalcocite zone at one place lies at a depth of 1,200 feet below the surface.

[4] W. Lindgren, personal observations; C. F. Tolman, *Min. and Sci. Press*, Nov. 6, 1909; W. H. Truesdell, *Min. and Sci. Press*, June 5, 1909; W. H. Weed, *Mining World*, Jan. 14, 1911.

covered and developed by churn drills. The Ray mines are situated in a basin on the upper part of a creek, at an elevation of about 2,200 feet. The deposits are in an area of crushed and altered pre-Cambrian schist, cut by dikes of granite porphyry and diabase. The upper leached zone, containing some oxidized copper ore, is from 50 to 150 feet thick. The chalcocite, disclosed by drilling and underground operations, extends over a large area, probably more than 100 acres; its thickness is from 20 to 300 feet and in a considerable part of the area averages 60 feet. The chalcocite zone is richest at the top and gradually becomes poorer in depth. An ore-body of about 83,000,000 tons, said to average 2.2 per cent. copper, has been shown to exist, and the exploitation of this great deposit began in 1910. The region had long been known as copper-bearing and futile operations on small masses of oxidized ore along diabase dikes had been undertaken. Until about ten years ago, however, it was not thought that ore of so low a grade could be mined profitably. Water begins to come in at the lower limits of the ore-body, which lies below the level of the creek.

According to Weed the granite porphyry is later than the diabase, and it is probable that both at Miami and at Ray the intrusion of the granite porphyry was the cause of the primary mineralization of lean cupriferous pyrite in the schists. According to the same author there are some exceptions to the rule of sharp definition between the leached zone and the ore, for at the Ray Central mine the change is gradual and for some distance above the 200-foot level the ore consists of one-half chalcocite and one-half cuprite. The passage into primary sulphides is usually effected within 50 feet. In this deepest zone are found disseminated pyrite, a little chalcopyrite and molybdenite, and quartz veinlets. This material contains less than 1 per cent. copper.

Chuquicamata.—The great copper lode at Chuquicamata in which the developed ore is estimated to be over 300,000,000 tons carrying an average of a little more than 2 per cent. copper, is situated in Northern Chile. The climate is exceedingly arid. The mass of the ore so far developed is oxidized and carries mainly brochantite, near the surface also atacamite, the latter formed by aid of windblown sodium chloride. The primary ore as shown by borings and by adjacent smaller deposits contains mainly quartz, pyrite and enargite. The present great oxidized body is formed by the oxidation of a deep chalcocite-covellite zone of which the lower part only now remains. This is in part mixed with oxidized ore and these ores are somewhat richer than those of the oxidized zone. During the chalcocitization the arsenic was removed and the oxidized zone now contains only traces of that element. During oxidation in the nearly rainless climate very little copper has been carried downward. The brochantite which contains nearly as much copper (70 per cent.) as chalcocite (80 per cent.) has evidently simply replaced the supergene sulphides. Further alteration of brochantite results in the poorer sodium-copper sulphates, kröhnkite and natrochalcite. Close to the surface is an irregular and shallow zone of leaching, in which the copper is partly removed and basic iron sulphates, hematite and gypsum have formed.

ZINC

Minerals.—The primary ore minerals of zinc are few in number. In fact sphalerite or zinc blende (ZnS) may be said to be the only one but that is almost universally present in sulphide ores and is a persistent mineral ranging from deposits of magmatic origin to deposits formed practically at the surface where reducing conditions obtain. Wurtzite, the hexagonal modification of zinc sulphide is questionable as to its hypogene origin. The absence of zinc sulpharsenides and sulphantimonides is remarkable and the small quantities of zinc reported in analyses of such minerals may well be caused by mechanically admixed zinc sulphide. Zinc spinel or gahnite $ZnO.Al_2O_3$) occasionally occurring in high temperature deposits is of no economic importance. The minerals zincite (ZnO), franklinite ($ZnO.Fe_2O_3$), willemite (Zn_2SiO_4), troostite ($Zn(Mn)_2SiO_4$), and several rare silicates are almost exclusively confined to the unique deposits at Franklin Furnace, New Jersey.

The oxidized ores of supergene origin comprise the m common smithsonite ($ZnCO_3$), the calamine (ZnH_2SiO_5) and the hydrozincite $ZnCO_3.2ZnO_2H_2$). Willemite may apparently also be formed during oxidation. There is also the rarer aurichalcite, a basic carbonate of copper and zinc, and several still more infrequent arsenates and vanadates. Goslarite ($ZnSO_4.7H_2O$) forms efflorescences but is of principal interest as the form in which zinc is usually transported in solution.

Monheimite (ZnFe) CO_3, the iron rich variety of smithsonite, is not uncommon.

Solubility and Mineral Development.—The sulphate and chloride of zinc are very easily soluble, whereas the carbonate and the silicates are difficultly soluble. Zinc blende is attacked by oxygenated water and the carbonate forms slowly. In sulphuric acid zinc blende is fairly easily soluble with development of H_2S and hence the oxidation of the mineral proceeds most rapidly in pyritic deposits. Unless limestone or some other precipitant is available the zinc of the oxidized zone is rapidly dispersed as sulphate and many examples are known of zinc-bearing sulphide deposits from the oxidized part of which the metal has wholly disappeared. Zinc is, in fact, the most mobile of the common metals in ore deposits. Like galena, zinc blende is also readily attacked by ferric sulphate:

$$ZnS + 4Fe_2(SO_4)_3 + 4H_2O = ZnSO_4 + 8FeSO_4 + 4H_2SO_4$$

and the resulting free acid starts the decomposition again.

Smithsonite is supposed to form according to the reaction $ZnSO_4 + CaCO_3 = ZnCO_3 + CaSO_4$ but there is reason to believe that where the replacement is effected by equal volumes the dilute solution of zinc carbonate is rather the reagent than the sulphate. It is not uncommon to find limestone replaced by smithsonite with perfect preservation of structure. The oxidized zinc ores are often inconspicuous earthy or admixed with clay and ferric hydroxide, and, therefore, easily escape attention. Zinc sulphate is often found in mine waters.

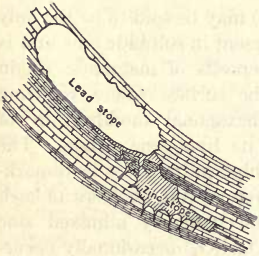

Fig. 282.—Diagram illustrating development of oxidized zinc ore in limestone below primary bodies of lead-zinc ore, May Day mine, Tintic, Utah. *After G. F. Loughlin.*

Supergene Shoots of Zinc Ore.—In calcareous rocks the descending zinc solutions are easily arrested and there it is common to find secondary zinc shoots below the primary ore (Fig. 282). In case of the extremely common combination of zinc and lead the latter metal remains in its original place as residual galena or cerussite while the smithsonite is found lower down or along convenient paths used by the downward moving waters. G. F. Loughlin[1] has described many cases of this kind from Tintic, Utah, and other places. Usually smithsonite forms first and hydrozincite, calamine and aurichalcite are distinctly secondary after smithsonite.

Supergene Zinc Sulphide.—Zinc is not as a rule deposited as a secondary (supergene) sulphide and no case has been recorded where it replaces pyrite, as chalcocite so often does. Many cases have, however, been described showing that zinc sulphide may form below the oxidized zone. W. H. Weed describes such an occurrence at Neihart, Montana, and F. Bain regards a certain red variety of zinc blende at Joplin, Missouri, as of secondary origin (p. 451). White amorphous zinc sulphate has been found precipitated by hydrogen sulphide from mine waters. Crystals of zinc blende have been observed in old workings opened after having been flooded for many years.[2]

Wurtzite, the hexagonal form of zinc sulphide, which, except by its optical qualities, is almost indistinguishable from sphalerite has been lately discovered at several places in the United States, particularly at Joplin, at Butte, at the Hornsilver Mine, Utah, at Goldfield, Nevada, and at the Era district, Idaho.[3]

Allen and Crenshaw[4] have shown a remarkable analogy between pyrite and marcasite on one hand and sphalerite and wurtzite on the other. While pyrite and sphalerite may crystallize from alkaline or from acid solutions the presence of free acid is essential for the formation of marcasite and wurtzite. It is believed that either wurtzite develops from sphalerite under the influence of acidic solutions or that both crystallize together from such solutions. Butler believed that the wurtzite which has formed abundantly in the lower levels of the oxidized zone at the Hornsilver mine[5] is of secondary origin. The oxidation of sulphides yielded sulphates and free acid. The difficultly soluble copper sulphide was precipitated as covellite on the more easily soluble

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[1] *Econ. Geol.*, vol. 9, 1914, p. 1.

[2] C. R. Keyes, *Trans.*, Am. Inst. Min. Eng., vol. 31, 1901, p. 611.
W. P. Jenney, *idem*, vol. 33, 1903, p. 470.

[3] For a review of these occurrences see J. B. Umpleby, *Prof. Paper* 97, U. S. Geol. Survey, 1917, pp. 87–89.

[4] *Am. Jour. Sci.*, 4th ser., vol. 34, 1912, pp. 341–396; *idem*, vol. 38, 1914, pp. 393–431.

[5] B. S. Butler, *Prof. Paper* 80, U. S. Geol. Survey, 1913, p. 154.

sphalerite, which, however, could not precipitate wurtzite, this being the more soluble of the two zinc sulphides. After the acid had become partly neutralized, the wurtzite would be precipitated by the hydrogen sulphide generated by the attack of H_2SO_4 on the sphalerite. Occurrences of "Schalen blende" or concentric intergrowths of sphalerite and wurtzite are known from Vieille Montagne, in Belgium, and other places. There is, as yet, much uncertainty as to the conditions under which wurtzite develops in nature.

LEAD

Minerals.—Among the primary lead minerals galena (PbS) is by far the most common; bournonite ($PbCuSbS_3$), jamesonite ($Pb_2Sb_2S_5$) and a host of other lead sulphantimonides such as boulangerite, geocronite, etc., are of little economic importance.

Far more numerous are the oxidized lead ores. They comprise the red oxide, minium (Pb_3O_4), the yellow massicot (PbO) and the dark plattnerite (PbO_2). All these are comparatively rare as are the chloride and the oxy-chloride. The really abundant oxidized lead minerals are anglesite ($PbSO_4$) and especially cerussite ($PbCO_3$). Of considerable importance are also a series of hydrous basic sulphates of lead with copper and iron, including the blue linarite $(Pb,Cu)SO_4(Pb,Cu)(OH)_2$, the yellow plumbojarosite ($PbO.3Fe_2O_3.4SO_3.6H_2O$), the microscopic foils of which cling to the finger like graphite and several other yellow earthy lead copper sulphates of varying composition.

Lead chloro-phosphate called pyromorphite ($Pb_5P_3O_{12}Cl$) and the corresponding arsenate (mimetite) and vanadate (vanadinite) are not uncommon and may be considered ore minerals. The same applies to wulfenite ($PbMoO_4$), and crocoite ($PbCrO_4$) and stolzite ($PbWO_4$).

Silicates of lead with several other oxy-salts like antimonio-silicate of manganese are found sparingly at the two abnormal deposits of contact-metamorphic type, Franklin Furnace, New Jersey and Långban, Sweden.

Reactions in the Oxidized Zone.—Lead in contrast to zinc shows slight mobility in the oxidized zone. All the salts are difficultly soluble, particularly the carbonate. The sulphate is very slightly soluble. Most soluble is the chloride[1] by means of which some transportation may be effected. Galena is slightly

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[1] At 15° C., 0.909 grams in 100 grams H_2O; at 100° C., 3.340 grams.

attacked by dilute H_2SO_4 and especially by the same solvent together with ferric sulphate.[1] The first change in galena is usually to anglesite ($PbS+4O=PbSO_4$). Residual nodules of galena are surrounded by dark concentric rings of anglesite, the color being caused by remaining, finely disseminated lead sulphide (Fig. 283). Anglesite is also seen well crystallized. Cerussite appears to form easily from anglesite and usually predominates ($PbSO_4+H_2CO_3=PbCO_3+H_2SO_4$); it appears as beautiful crystal groups but is more commonly earthy, white or yellowish and of sandy texture (sand carbonate).

FIG. 283.—Photomicrograph of polished section showing residual galena (light) in anglesite (dark). Magnified 420 diameters.

Once formed these two minerals are exceedingly stable. However, if there is free sulphuric acid or ferric sulphate or chlorides present the lead may be rendered more mobile.[2] Soluble chloride and oxy-chlorides form and a whole series of basic yellow lead iron sulphates may be developed. Plumbojarosite, one of this

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[1] H. C. Cooke, *Econ. Geol.*, vol. 21, 1913, p. 11.

[2] One liter of water dissolves only 4.4 milligrams $PbSO_4$ while the same amount of saturated NaCl solution dissolves 660 milligrams, slowly decomposing it to chloride.

series, has been used as an ore.[1] Considerable migration of lead in oxidized ore has been carried on by the aid of the two reagents just referred to.

"Steel galena" owes its fine-grained texture either to mechanical deformation of larger crystals or to a beginning transformation to anglesite. It is supposed to be rich in silver but this is by no means always true. Oxidized lead ores are usually poor in silver, but here again there may be many exceptions noted.

Wulfenite is formed from the oxidation of galena and molybdenite; the latter is often present only in microscopic particles. The enrichment of lead in the oxidized zone is generally a consequence of the solution of associated minerals and the resulting reduction of volume of the ore.

Supergene Sulphides.—Well-defined zones of supergene lead sulphide have never been observed. According to its position in Schürmann's series lead should be deposited as sulphide ore on sphalerite and pyrite.

Examples are known of galena formed as well-defined crystals on iron spikes from old workings of a lead mine in Missouri. Thin films of galena are sometimes deposited on zinc blende; this has been observed by Irving and Bancroft at Lake City, Colorado; by Boutwell at Bingham, Utah; and by Ransome at Breckenridge, Colorado.

Oxidation in the Coeur d'Alene District.—In the Coeur d' Alene district,[2] northern Idaho, the precipitation is heavy, the topography is accentuated, and the water level stands near the surface of the ground. The veins, which are enclosed in quartzite country rock, contain galena, zinc blende, and siderite. Little pyrite is present. The lower limit of oxidation is very irregular. Complete oxidation is confined to the vicinity of the surface, but cerussite occurs in vugs and fractures several hundred feet below the surface, while the galena may in places occur in the outcrops. The minerals first attacked are pyrite and sphalerite, while the solid galena is very resistant. The chief products of oxidation are limonite, occurring as great masses at the outcrops, and cerussite, to which as the latest products pyromorphite, also plattnerite (PbO_2) are added. Owing to the

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[1] B. S. Butler, *Prof. Paper* 80, U. S. Geol. Survey, 1913, p. 109.

[2] F. L. Ransome and F. C. Calkins, *Prof. Paper* 62, U. S. Geol. Survey, 1908, p. 132.

prevalence of siderite, cerussite is the predominating oxidized lead mineral; anglesite is absent. The quantity of silver contained in the galena is small and there is no evidence of enrichment either in the oxidized zone or below it.

Oxidation in the Mississippi Valley District.—The effects of oxidation on the lead and zinc deposits of the Ozark region have been described by Bain,[1] Siebenthal and Smith,[2] and Buckley and Buehler.[3] The oxidation is mostly confined to the zone between the surface and the water level, which is rarely as much as 100 feet below the surface. The oxidized ore is, to a considerable degree, a product of the residual weathering of limestone and chert and thus consists of a confused mass of red residual clay, with layers and fragments of white chert, in which are found galena and the oxidized ores of lead and zinc. Galena is the only sulphide found in quantity above the water level.

The zinc blende alters either to calamine or to smithsonite and nodular masses of each sometimes hold a kernel of the sulphide. A soft "tallow clay," mainly an impure kaolin with admixed calamine, occurs in many deposits. Bain asserts that secondary sulphides of zinc and lead with pyrite and chalcocite have been deposited at Joplin, below the zone of oxidation. Possibly this has taken place on a small scale, but most of the ore immediately below the oxidized zone appears to be of primary origin. Zinc blende and galena may easily have been deposited by the reduction of sulphate solutions by means of metallic iron (spikes), as illustrated in Bain's report, or by organic matter. There is scarcely enough pyrite in the Missouri deposits to cause extensive replacement (according to Schürmann's reactions) of pyrite by galena and zinc blende. Moreover, should the deposition have taken place by precipitation on sulphides, galena could replace zinc blende, but the reverse reaction could not take place. There are undoubtedly two generations of blende, one earlier in the chert and another and less important deposition of small red crystals on dolomite, but this scarcely proves that enrichment has taken place. During oxidation the solutions are probably

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[1] H. Foster Bain, Lead and zinc deposits of the Ozark region, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, pp. 155–162.

[2] C. E. Siebenthal and W. S. Tangier Smith, Joplin folio, *Geol. Atlas* 148, U. S. Geol. Survey, 1907.

[3] E. R. Buckley and H. A. Buehler, Geology of the Granby area, *Missouri Bur. of Geology and Mines*, vol. 4, 2d ser., 1906.

acid only where considerable amounts of iron sulphide are present. A frequently occurring association is that of calamine surmounted by crystals of dolomite, which could not have been deposited from solutions containing free sulphuric acid.

In the upper Mississippi Valley (p. 457), according to C. R. Van Hise,[1] the main valuable minerals are smithsonite and galena; above the level of ground water, which lies close to the surface. Encrusting the galena are some cerussite and less anglesite; with the smithsonite is some zinc blende. The smithsonite may extend 15 to 30 feet below water level, but at greater depth the oxidized products almost wholly disappear. Below this level zinc blende with much marcasite forms the principal ore-bodies. Above the water level, then, the iron sulphide has been dissolved, as well as much of the zinc blende, leaving a richer concentrate of galena. The galena at or below the water level may possibly in part be secondary, precipitated by a reaction between lead sulphate and iron sulphide. At a still lower level secondary zinc blende may have been deposited, but no cogent proof of this has been furnished.

GOLD

Gold shows slight mobility in ore deposits; it is less easily transported than silver, and, compared to copper and zinc, it is almost stationary. The solubility of gold has been discussed briefly in the chapter on placers. Gold usually occurs native, in coarse or in fine distribution.

The tellurides form the only definitely known combinations of gold with other elements. Among them are calaverite and the less common sylvanite and krennerite, all with the general formula of $(Au, Ag) Te_2$, and petzite $(Au, Ag)_2 Te$. The tellurides are apparently able to form under widely differing conditions, though they are generally absent from the deposits formed under conditions of very high pressure and temperature. They decompose easily above the water level; the tellurium is in part carried away as soluble compounds, in part fixed as tellurite (TeO_2) or tellurates of iron like emmonsite and durdenite. The gold remains in minute brownish grains (mustard gold). In most cases there is little evidence of solution and transportation of this gold.

Certain deposits formed by hot ascending waters near the

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[1] *Trans.*, Am. Inst. Min. Eng., vol. 30, 1900, pp. 102–109.

surface contain selenium, either alone or together with tellurium (Republic, Washington; Tonopah, Nevada; Radjang Lebong, Sumatra), and probably they carry a selenide of gold, though its existence has not been definitely proved. Little is known about the oxidation products of selenium.

Much more commonly the gold occurs in native form, in gold-quartz veins with small amounts of sulphides. In such deposits there is little evidence of solution and transportation of the gold. Enrichment often takes place in them, but rather by reduction of volume of the ore than by solution of gold. The sulphides in these veins are usually rich in finely distributed gold which remains behind upon oxidation. The oxidation of a crystal of pyrite will generally result in a pseudomorph of limonite which contains flakes of native gold, indicating that within the crystal a certain amount of transportation of gold has taken place.

Very common also are the deposits in which the sulphides abound and which contain no visible free gold. The Gilpin County veins, Colorado; Mount Morgan, Queensland; and the Haile Deposit, South Carolina, may serve as examples. It is in these deposits that most evidence is found of the solution and transportation of gold.

Gold has undoubtedly been transported in the form of chloride, though its migration in colloidal suspension derived by solution processes from native gold is probably much more common than has been suspected. Now, while chloride of gold is easily soluble in water, it is also most easily precipitated by reducing reagents, such as organic matter, ferrous sulphate, metals or sulphides, like pyrite. In the older literature ferric chloride and ferric sulphate were frequently given as solvents of gold, and the question had previously been discussed by Pearce, Don, and T. A. Rickard. H. N. Stokes[1] and J. R. Don[2] showed that ferric sulphate is ineffective, and the recent work of W. H. Emmons[3] and A. D. Brokaw[4] has shown that gold is insoluble in ferric chloride also and goes into solution only when nascent chlorine

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[1] *Econ. Geol.*, vol. 1, 1906, p. 650.

[2] *Trans.*, Am. Inst. Min. Eng., vol. 27, 1898, p. 599.

[3] The agency of manganese in the superficial alteration and secondary enrichment of gold deposits in the United States, *Trans.*, Am. Inst. Min. Eng., vol. 40, 1910, pp. 767–837.

[4] *Jour. Geology*, vol. 18, 1910, p. 322; vol. 21, 1913, pp. 251–267.

is present. The metal is easily precipitated from chloride solution, most easily and completely by ferrous sulphate.

In deposits containing much pyrite oxidation results in the liberation of sulphuric acid. Sodium chloride is present in some degree in all mine waters and is abundant in some. Reaction between sulphuric acid and sodium chloride results in hydrochloric acid, and it should be recalled that Don actually found free HCl in a number of superficial mine waters from New Zealand.[1] If dioxide of manganese is present in the deposit nascent chlorine will be generated according to the formula:

$$MnO_2+4HCl=2H_2O+MnCl_2+2Cl.$$

Ferric and cupric salts have similar power, but chlorine develops very slowly, if at all, in the cold. It should be expected, according to Emmons, that auriferous deposits which contain manganese would show the effect of solution and migration of gold more clearly than non-manganiferous ores. According to experiments by A. D. Brokaw, quoted by Emmons, gold is not dissolved in hydrochloric acid, ferric sulphate, or ferric chloride.[2] It is dissolved at 38° C. in concentrated solution containing both ferric sulphate and hydrochloric acid; also at the same temperature in concentrated solution of cupric chloride and hydrochloric acid; the dilute solutions are not effective. Brokaw's experiments verified the solubility of gold by nascent chlorine in presence of manganese as outlined above. It was shown that a small piece of rolled gold weighing about 0.15 gram in a solution of 50 c.c. of one-tenth normal HCl with 1 gram of powdered MnO_2 in 14 days lost 0.01369 gram. To approximate natural waters a solution one-tenth normal was made as to ferric sulphate and sulphuric acid and one twenty-fifth normal as to sodium chloride. In a second experiment 1 gram of MnO_2 was added; the time allowed was 14 days:

$$Fe_2(SO_4)_3+H_2SO_4+NaCl+Au.$$

No weighable loss.

$$Fe_2(SO_4)_3+H_2SO_4+NaCl+Au+MnO_2.$$

Loss of gold, 0.00505 gram.

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[1] J. R. Don, *op. cit.*

[2] Don in 1897 had already stated that, in the absence of free chlorine, gold is insoluble both in $FeCl_3$ and $Fe_2(SO_4)_3$.

Where much MnO_2 is present, ferrous sulphate is almost immediately transformed to ferric sulphate and the precipitation of the gold is delayed.[1]

The gold dissolved in the presence of MnO_2 and held in solution by the absence of $FeSO_4$ moves downward until the excess of acid is reduced, and simultaneously the iron and manganese compounds tend to hydrolyze and deposit oxides. At this stage $FeSO_4$ becomes increasingly prominent and effective as a precipitant. The transportation of gold is thus dependent upon the oxidation of ferrous sulphate by manganese dioxide. In the presence of MnO_2 gold may even be carried down and deposited below the water level.

The greater enrichment in gold will be found in the lower part of the oxidized zone. A slight enrichment is sometimes found in the upper part of chalcocite zones. In zones of deep oxidation as at Tintic,[2] Utah, supergene concentrations of gold may be found at all levels. Manganese is apparently always present here. The supergene gold is usually very pure.

Many examples of actual redeposition of gold are mentioned in the literature, such as films of gold in fissures and on other minerals—for instance, on zinc blende at mines in Lake City, Colorado.[3]

Crystals of gold are sometimes deposited on the surface of pyrite crystals; fine examples of this have been described from the Crystal mine, Port Snettisham, Alaska.[4] Associated with the gold were small crystals of galena. This is, however, probably a primary deposition from alkaline ascending solutions.

Organic matter, hydrogen sulphide, carbon, sulphides, tellurides and carbonates[5] precipitate gold from chloride solution. Palmer and Bastin[6] have shown that most sulphides easily precipitate gold; pyrite and galena, which do not precipitate silver readily bring down the gold. With dilute solutions a protective colloid may inhibit precipitation.[7]

A peculiar feature in certain gold deposits where extensive

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[1] W. H. Emmons, *op. cit.* (Experiment 20.)

[2] W. Lindgren, *Econ. Geol.*, vol. 10, 1915, p. 237.

[3] J. D. Irving, oral communication.

[4] J. E. Pogue, *Zeitschr. Kryst. u. Min.*, vol. 49, 1911, p. 455.

[5] V. Lenher, *Econ. Geol.*, vol. 13, 1918, pp. 161–184.

[6] *Econ. Geol.*, vol. 8, 1913, pp. 140–170.

[7] E. S. Bastin, *Jour.* Washington Acad. Sci., 1916, p. 64.

kaolinization has taken place near the surface is the occurrence of white kaolin extraordinarily rich in gold so fine that it is scarcely visible when the material is washed in the pan.[1] This is undoubtedly an effect of oxidation, but the mode of this enrichment is not fully explained. Possibly the gold has been precipitated in this extremely finely divided form from colloidal solutions.

Examples of Oxidation of Gold Deposits.—In the Blue Mountains of Oregon,[2] a region of heavy precipitation but dry summers and rather high topographic relief, gold-quartz veins are contained in Paleozoic argillites and in intrusive diorite. The ores, which in places carry much free gold, are oxidized down to a depth of 100 to 300 feet. At the Sanger mine, on Eagle Creek, the uppermost 100 feet showed a narrow vein, caused by collapse of outcrops, yielding $25 per ton, while farther down the vein widened and its gold was reduced to $12 per ton.

In the highly productive gold veins of the Cracker Creek and Granite districts the sulphides and arsenopyrite are in fine distribution and much of the ore is rather hard. The water level is high, but on the steep hillsides the oxidized zone is in places 250 feet deep. The oxidation to this depth is only partial, but there is a surprisingly slight difference in tenor between the surface ore and the primary ore. In the latter the gold is contained mainly in the sulphides; free gold is present in the oxidized ore, but there is not enough to convert the material into free-milling ore. No great reduction of volume has taken place, and weathering has only slightly increased the tenor of gold, while the small silver content has been slightly leached.

The gold-telluride lodes of Cripple Creek, Colorado,[3] are mainly sheeted zones in which the seams are filled with quartz, fluorite, and calaverite ($Au(Ag)Te_2$). These deposits oxidize to brownish clayey material in which the original vein structure is no longer apparent. As quartz is not abundant, the main product of the oxidation is kaolin, with some limonite. The fluorite is carried away, while the tellurides are very easily reduced to dark-brown

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[1] W. Lindgren, *Twentieth Ann. Rept.*, U. S. Geol. Survey, pt. 3, 1900, p. 171. F. Guitermann, *Proc.*, Colorado Sci. Soc., vol. 3, 1891, pp. 264–268.

[2] W. Lindgren, The gold belt of the Blue Mountains of Oregon, *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901.

[3] W. Lindgren and F. L. Ransome, *Prof. Paper* 54, U. S. Geol. Survey, 1906.

powdery gold. The tellurium is partly carried away in solution but to some extent remains as colorless tellurite (TeO_2) and green ferric tellurites like durdenite and emmonsite.

The oxidation extends to the water level, which is from 300 to 900 feet below the surface, and in places the ore is oxidized for some distance below the water level. There has been little or no enrichment of gold in the oxidized zone, but a decided leaching of the small amount of primary silver originally contained as telluride or tetrahedrite. No secondary silver sulphides were detected, nor is there evidence of secondary deposition of tellurides.

At Creede,[1] Colorado, and in the Tomboy and the Camp Bird mines,[1] Colorado, as well as in the Granite-Bimetallic[1] mine, Montana, in the Tintic mines,[2] Utah, and in the Mount Morgan mine, Queensland (p. 861) examples are found of the deposition of supergene gold in the lower part of the oxidized zone.

SILVER

Minerals.—Of the primary silver minerals argentite (Ag_2S) is easily the most important. Hessite (Ag_2Te) is not uncommon in gold-quartz veins but plays no great part as an ore mineral, and the selenide is a rarity. Native silver may under some circumstances—as in the Lake Superior copper mines—appear as a primary mineral. Of more importance are tetrahedrite ($Cu_8Sb_2S_7$) and tennantite ($Cu_8As_2S_7$) for they usually contain silver sulphide in chemical combination. Finally we have a long series of sulphantimonides, sulpharsenides and sulphobismuthides of silver or silver-lead. All of these are invariable of late though not necessarily of supergene origin. At least two of them—proustite and pyrargyrite—are probably often of hypogene derivation.

Galena and silver are particularly frequently associated and it is now known that this silver is contained in galena as minute inclusions of argentite[3] easily made visible by etching a polished surface with HCl. Occasionally tetrahedrite is also included in galena. It is probable that the argentite is also contained in other of the common sulphides and sulpho salts. Galena and argentite form a eutectic at 77 per cent. Ag_2S at 630° C. but this is not developed where there is less than 2.70 per cent.

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[1] W. H. Emmons, *Bull.* 625, U. S. Geol. Survey, 1917, pp. 324–349.

[2] W. Lindgren, *Econ. Geol.*, vol. 10, 1915, p. 237.

[3] A. E. Nissen and S. L. Hoyt, *Econ. Geol.*, vol. 10, 1915, pp. 172–179.

Ag_2S and it has not been observed in nature. Silver sulphide may exist in solid solution in PbS but the limit of this lies below 0.2 per cent. Ag_2S.

The distinctly supergene silver minerals comprise native silver, argentite, cerargyrite ($AgCl$), embolite $Ag(ClBr)$, bromyrite ($AgBr$) and iodyrite (AgI). The first four are of common occurrence.

Stromeyerite ($CuAgS$,) closely allied to chalcocite, is probably always of supergene origin and occurs in chalcocite zones, for instance, at Broken Hill, New South Wales.

Of the group of sulpho salts, referred to above, the following are surely supergene though they may well also develop as a last effect of hypogene action, perhaps by an "upward secondary enrichment:"

Pyrargyrite	Ag_3SbS_3	Proustite	Ag_3AsS_3
Miargyrite	$Ag\ SbS_2$		
Stephanite	Ag_5SbS_4		
Polybasite	Ag_9SbS_6	Pearceite	Ag_9AsS_6

Dyscrasite is really an alloy of silver and antimony with varying composition; it has formed important ore at Broken Hill and Chañarcillo and is also known from Cobalt.

Solubility and Mineral Development.—Argentite is oxidized to cerargyrite (horn silver), probably by way of the fairly soluble silver sulphate, as follows: $Ag_2S+4O=Ag_2SO_4$; $Ag_2SO_4+2NaCl=2AgCl+Na_2SO_4$. Pseudomorphs of cerargyrite after argentite are well known.

Argentite is very slightly attacked by dilute sulphuric acid, but is much more rapidly decomposed by ferric sulphate or by a mixture of the two solvents.[1] Pyrargyrite and polybasite are also decomposed by sulphuric acid, silver and a little antimony going in solution. The reaction is increased by ferric sulphate.

It is thus evident that the presence of oxidizing pyrite will greatly facilitate the movement of silver in ore deposits and while the metal is not as mobile as copper or zinc it is transported much easier than gold. The universal presence of chlorine in waters, particularly in those of arid climates, and the insolubility of the silver chloride account for the common occurrence of the latter mineral in the oxidized zones of ore deposits.

Though both cerargyrite and argentite are easily reduced to

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[1] H. C. Cooke, *Jour. Geology*, vol. 21, 1913, p. 13.

native silver they are very slightly soluble in solutions usually occurring in nature.

The solubility of the various silver salts is as follows, in grams of anhydrous salt per 100 grams of water: Ag_2SO_4 at 25° C., 0.83; at 100° C., 1.46. Ag_2CO_3 at 25° C., 0.003; in water saturated with CO_2, at 15° C., 0.08. AgCl at 13° C., 0.00014; at 43° C., 0.0004. AgBr at 25° C., 0.00001. AgI is still less soluble than the bromide. It will be noted that the solubility of AgCl increases rapidly with the temperature.[1]

Another way to facilitate transportation of silver, which seems to have escaped general attention, is by colloid solutions or suspensions. Such solutions of AgCl, AgBr, AgI and Ag_2S have been prepared by methods which might easily be used by nature and by adding protective colloids like silica they may be made very stable. The possibilities of such transportation are evident.[2]

Precipitation.—Native silver is easily attacked by ferric sulphate; the silver salts are easily reduced by ferrous sulphate.[3]

The reaction with ferric sulphate is reversible silver being dissolved on heating and reprecipitated on cooling. It may be written as follows: $2Ag+Fe_2(SO_4)_3 \leftrightarrows Ag_2SO_4+2FeSO_4$.

Silver is further precipitated from its solutions by organic matter as well as by copper, cuprite and various sulphides.[4] The precipitation by sulphides and arsenides has been investigated by C. Palmer and E. S. Bastin[5] and also by F. F. Grout[6] who found that alabandite, chalcocite, covellite and enargite precipitate silver rapidly while pyrite, chalcopyrite, galena and sphalerite are comparatively inactive. All simple arsenides are also very effective precipitants while the sulpharsenides do not react.

Silver is further precipitated by many silicates such as kaolin

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[1] A. G. Melcher, The solubility of silver chloride, etc., *Jour.* Amer. Chem. Soc., vol. 32, 1910, pp. 50–56.

[2] See E. S. Bastin, Experiments with colloidal gold and silver, *Jour.* Washington Acad. Sci., 1916, p. 64.

[3] H. N. Stokes, *Econ. Geol.*, vol. 1, 1906, p. 649.

H. C. Cooke, *Jour. Geology*, vol. 21, 1913, pp. 1–28.

[4] J. H. L. Vogt, Ueber die Bildung des gediegen Silbers, *Zeitschr. prakt. Geol.*, 1899, p. 113.

[5] *Econ. Geol.*, vol. 8, 1913, pp. 140–170; C. Palmer, *Idem*, vol. 12, 1917, pp. 207–218.

[6] *Econ. Geol.*, vol. 8, 1913, pp. 407–433.

and orthoclase.[1] Calcite, siderite and rhodochrosite do not precipitate silver from weak solutions.[2]

Silver sulphide is precipitated by hydrogen sulphide which may result from the attack of H_2SO_4 on pyrrhotite, zinc blende or galena.

Supergene Sulphide Enrichment.—According to Schürmann's table (p. 843) silver sulphide should replace many other sulphides such as those of copper, lead, zinc and iron, following the reaction $Ag_2SO_4 + ZnS = Ag_2S + ZnSO_4$. Such replacements may be found but they are rare and the sulphides generally precipitate native silver (in places with some AgS) and in the case of galena, sphalerite and pyrite the reaction is weak. On the other hand almost all of the common hypogene sulphides as well as tetrahedrite, tennantite and enargite are easily replaced by the silver sulphantimonides,[3] such as pyrargyrite and to this action much of the supergene sulphide enrichment is due. These rich silver minerals as well as secondary argentite are also found in druses, veinlets and crusts in which form they do not represent replacements but are perhaps precipitated by hydrogen sulphide or by complex reactions between solutions.

According to the investigations of F. F. Grout[4] and L. G. Ravicz[5] the silver sulphantimonides and the less common sulpharsenides are formed where the descending solutions have become alkaline through the neutralizing reactions going on all the time during their downward travel.[6] Grout obtained a product of the approximate composition of stephanite by adding silver sulphate to a solution obtained by digesting stibnite in a 1 per cent. solution of sodium carbonate. At higher temperatures of 250° C. and 300° C. de Senarmont and Doelter obtained proustite, pyrargyrite and stephanite by treating alkaline sulpharsenite or sulphantimonite with alkaline carbonates.

It is a difficult task to separate sharply the effects of supergene alkaline waters from those of the ascending hypogene solutions. Both may yield argentite and sulpho salts. In all silver deposits there is a progressive hypogene series beginning

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[1] E. C. Sullivan, *Bull.* 312, U. S. Geol. Survey, 1907, pp. 37–64.

[2] L. G. Ravicz, *Econ. Geol.*, vol. 10, 1915, pp. 368–389.

[3] F. N. Guild, *Econ. Geol.*, vol. 12, 1917, pp. 297–352.

[4] *Econ. Geol.*, vol. 8, 1913, pp. 407–433.

[5] *Econ. Geol.*, vol. 10, 1915, pp. 378–484.

[6] G. S. Nishihara, The rate of reduction of acidity of descending waters, etc., *Econ. Geol.*, vol. 9, 1914, pp. 743–757.

with arsenopyrite, pyrite, sphalerite, tetrahedrite and galena, the later replacing the earlier; then follows a series of richer silver minerals such as pyrargyrite and proustite, which may replace any and all of the earlier sulphides[1] besides filling druses and veinlets. These may be either of supergene or hypogene origin.

Zones of Supergene Deposition.—Zones of the regularity of those of copper deposits are rarely found in case of silver ores. Still there is a more or less marked succession, dependent on the association of minerals. Silver enrichment apparently necessitates ferric sulphate as an active solvent; this may be furnished by pyrite or by arsenopyrite. In the gossan and down to the water level we find mainly cerargyrite sometimes as large masses, though even here there may be local accumulations of native silver and argentite. At Lake Valley, New Mexico,[2] 100 tons of silver was obtained from a comparatively small stope, called the Bridal Chamber; it was contained in Paleozoic limestone, and within 100 feet of the surface. Below the chloride zone we may find the chlorobromides and below this a horizon where the iodyrite is precipitated.[3] It would be expected to find the most difficultly soluble salt at the top instead of at the bottom but according to W. H. Emmons[4] the explanation is that if in the solution containing the three halogens chlorine is in vast excess silver chloride though most soluble will be precipitated first. The three halogens are probably of wind blown origin, from the salt pans of the desert or from the sea.

Below the chloride zone we find usually more native silver and dyscrasite and this may extend to considerable depths, in fact many hundred feet below the water level. The deposition of this native silver may, like that of argentite and sulpho salts, be accompanied by calcite, barite and other gangue minerals usually deposited by the ascending waters. Theoretically the zone of the argentite and the sulpho salts should begin at the water level but in silver deposits this transition line is far less sharply marked than in copper deposits. The bulk of the supergene argentite usually lies above the sulpho salts.

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[1] F. N. Guild, *Econ. Geol.*, vol. 12, 1917, pp. 297–352.

[2] Ellis Clark, *Trans., Am. Inst. Min. Eng.*, vol. 24, p. 148, 1895.

[3] F. A. Moesta, Ueber das Vorkommen der Chlor-Brom-und Jodver bindungen in der natur, Marburg, 1870, pp. 57.

J. A. Burgess, *Econ. Geol.*, vol. 6, 1911, p. 13; *Idem.*, vol. 12, 1917, p. 590.

[4] *Bull.* 625, U. S. Geol. Survey, 1917, p. 257.

The secondary silver sulphides may descend to considerable depths, in places 500 to 1,000 feet below the surface. Poorer ore of hypogene origin will be found below the enriched zone (Fig. 284). In many cases the oxidized ore and the supergene sulphides appear inextricably mixed.

Enrichment at Granite-Bimetallic Mine.—The Granite Bimetallic vein, in Montana, described by W. H. Emmons,[1] is contained in granite and has been opened to a depth of 2,600 feet. The early production amounted to $32,000,000 in gold and silver. The strong, steep fissure has a filling from 1 to 20 feet wide, of simple structure and remarkable horizontal and vertical persistence. The primary ore consists of quartz and rhodochrosite with much pyrite, arsenopyrite, galena, and tetrahedrite, some galena and zinc blende, and specks of pyrargyrite. The ore is of comparatively low grade, containing 20 to 30 ounces of silver and about $2 in gold per ton (Fig. 284).

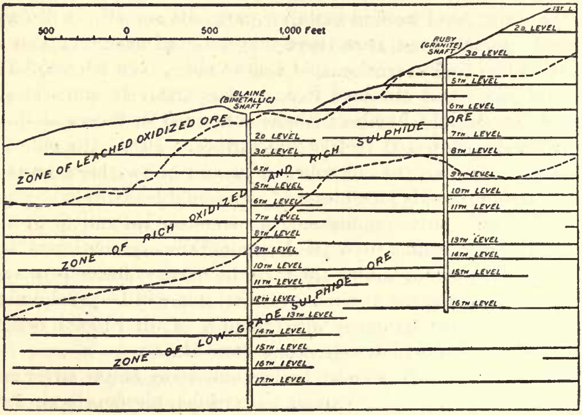

Fig. 284.—Longitudinal vertical projection of Granite-Bimetallic vein, Philipsburg, Montana, showing zone of enrichment. *After W. H. Emmons and F. C. Calkins.*

The uppermost, oxidized zone, from 50 to 300 feet deep, contains a poor iron-stained quartz, with little silver and a trace of gold. The material, which is so poor that many of the claims

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[1] *Prof. Paper* 78, U. .S Geol. Survey, 1913, p. 202.

along the vein were abandoned during its early history, also carries some cerussite, malachite, etc., as well as remnants of unoxidized sulphides. Emmons considers that this leached zone is derived by incomplete oxidation of the lower zone of enriched sulphides.

Below this leached zone lies the zone of enriched oxidized ore; it is for the most part between the 100-foot and 400-foot levels. This ore is composed of quartz, stained by iron and manganese, but contains much cerargyrite and native silver as thin seams. There is also some argentite, galena, etc. This ore contains much silver and from $5 to $16 in gold per ton. Its value is generally less than that of the ore in the next lower zone.

The zone of enriched sulphides extends in the main from 300 to 800 feet below the surface. It is extremely rich in silver and yields from $4 to $8 in gold per ton. Quartz and rhodochrosite are the gangue minerals; there is much argentite, pyrargyrite, and proustite, as well as abundant remains of primary ore minerals. The ruby silver, which is the most valuable mineral, is contained in veinlets or seams.

There is then here a distinct leached zone, a zone of moderate gold enrichment, and a zone of strong silver enrichment.

While it cannot be asserted that all of the primary ore was poor it is very unlikely that its value reached the high figures shown in the zone of secondary sulphides, which for a vertical distance of 300 to 400 feet averaged from 100 to 175 ounces of silver to the ton. Emmons concludes that the silver and gold from the upper parts of the vein, now eroded, have been carried down and that a successive enrichment has thus taken place. The moderate gold enrichment in the lower part of the oxidized zone is, according to Emmons, due to solution of gold by free chlorine (p. 880) and precipitation by ferrous sulphate.

Enrichment at Georgetown.—In the Georgetown district, Colorado,[1] argentiferous veins cut pre-Cambrian granites and schists. The climate is rigorous, the relief strong, the water level high. The zone of complete oxidation, in which the ores are rich in silver, extends at most 40 feet below the surface. Below this are friable black sulphides and secondary galena rich in silver and with more gold than occurs at greater depth; the primary sulphides, which contained about 25 ounces of silver per ton, are here enriched

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[1] J. E. Spurr and G. H. Garrey, *Prof. Paper* 63, U. S. Geol. Survey, 1908, p. 144.

and carry more than 200 ounces per ton. Below the zone where the soft secondary sulphides occur and irregularly overlapping the lower portion of this zone are rich ores containing polybasite, and ruby silver, better crystallized and more massive than the pulverulent sulphides, but also subsequent in origin to the primary galena-blende ore. These richer ores diminish in quantity as depth increases, although gradually and irregularly, so that the lower portion of the veins contains relatively less silver and lead. The best silver ore in most veins has been found in the uppermost 500 feet, although good ore extends locally down to 700 or 800 feet, and in one case 1,000 feet below the surface.

Enrichment at Tonopah.—At Tonopah, Nevada,[1] a series of rich silver-gold quartz veins (Au:Ag = 9:1 by weight), containing hypogene gold, argentite, polybasite, pyrite, etc., with rhodonite, adularia, and carbonates, cut across Tertiary andesites. The climate is exceedingly dry, and the veins are situated in a group of hills rising from the desert. The deposits are oxidized down to a depth of about 700 feet; ground water is lacking, but from 1,000 feet downward tepid and hot waters, containing mainly alkaline sulphates, are met. The oxidation is irregular and incomplete; the pyrite is changed to limonite, and much chloride of silver, with some bromide and iodide, has formed. In general, cerargyrite is found in the upper part of the oxidized zone; bromyrite in the middle, and iodyrite in the lower part. Other minerals of the oxidized zone are kaolin, hyalite, gypsum, limonite, hematite, jarosite, and wulfenite. On the whole the oxidized ore contains more silver than the primary ore.

Below the oxidized zone the fissures and cracks in the hard quartzose ore contain some secondary argentite, polybasite, pyrargyrite, and also chalcopyrite. Even the oxidized zone contains some veinlets filled with pyrargyrite. There is no well-defined zone of sulphide enrichment; secondary pyrite, blende, and galena are absent. Spurr believes that "the secondary sulphides in the oxidized zone originated from descending surface waters, and probably part but not all of the sulphides in druses in the sulphide ore have a similar origin." Manganese is present but no evidence of transportation of gold has been found.

Analyses show that the carbonates were removed from the

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[1] J. E. Spurr, *Prof. Paper* 42, U. S. Geol. Survey, 1905.

J. A. Burgess, *Econ. Geol.*, vol. 6, 1911, pp. 13–21.

E. S. Bastin and F. B. Laney, *Prof. Paper* 104, U. S. Geol. Survey, 1918.

primary ore and with them most of the lime and magnesia; iron, manganese, copper, lead, and zinc have been largely removed, likewise most of the selenium, arsenic, and antimony. The argentite has largely remained unaltered, while polybasite and the selenides have been decomposed; a little As and Sb remain to form ruby silver. The silver in the oxidized ore is combined as argentite, haloid compounds, and gold alloy.

Enrichment at Chañarcillo.—The exceedingly rich silver veins of Chañarcillo in the arid region of central Chile have been described by F. A. Moesta[1] and lately by W. L. Whitehead.[2] The veins intersect Mesozoic limestone with associated volcanics. At a depth of from 500 to 1,000 feet the primary ore appears. It consists of calcite and barite as the earliest minerals; a second stage is represented by pyrite, zinc blende, chalcopyrite and galena; a third stage by arsenopyrite and quartz; and a fourth stage in which tetrahedrite, pearceite, proustite, polybasite and pyrargyrite were deposited. The sulphide enrichment, in which processes of replacement predominate, occupies a vertical interval of 300 to 600 feet.

During the first stage of enrichment stephanite, pearceite, polybasite, stromeyerite and argentite replace earlier sulphides. In a later stage silver and dyscrasite develop on a large scale mainly by replacing sulpho salts and sulphides but also by replacing calcite and by filling.

The zone of oxidation is from 150 to 600 feet deep and is marked by the rich development of iron oxides, cerargyrite, embolite and iodyrite, the latter being mostly found in depth. The halides replace silver and dyscrasite, but also sulphides and calcite; to a minor extent they are deposited in open space. The succession is closed by a development of argentite and native silver in local enrichment due to a reversal of oxidation reactions; these minerals again replace the halides.

OTHER METALS

Platinum and Palladium.—The main source of platinum is the native metal in alloy with smaller amounts of others of the same group. Some sulphide deposits of the high temperature types contain sperrylite ($PtAs_2$) and usually also more palladium

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[1] *Op. cit.*

[2] *Econ. Geol.*, vol. 14, 1919, pp. 1–45.

than platinum; it is not known in what form the palladium appears. Platinum minerals are highly resistant to oxidation.

Sperrylite is not easily attacked by meteoric waters and this leads to concentration by reduction of volume. At Sudbury the shallow oxidized zone is richer in the platinum minerals than the primary ore.

The occurrence described on p. 791, from southern Nevada, shows that oxidized ores with much plumbojarosite (basic plumbo-ferric sulphate) contains finely divided gold as well as metallic platinum-palladium. It is probable that the metals have been precipitated from colloid solutions or suspensions.

Palladium with some platinum is found in the blister copper from certain smelters, especially from some treating ore from disseminated chalcocite deposits. The Ely deposits, Nevada (p. 865, yield more palladium than others. Selenium, common in all blister coppers, is here also present in unusual amounts. A. Eilers[1] showed that this copper contained in per cent. 0.0004 Pt, 0.00016 Pd and 0.055 Se but that there is no tellurium. This suggests a concentration of the platinum metals, possibly as selenides, during the supergene chalcocite enrichment.

Mercury.[2]—Cinnabar, the principal ore mineral of mercury, is stable up to its sublimation point (680° C.). It is practically insoluble in water (p. 838), but is soluble in alkaline sulphides forming the compounds $HgS.2Na_2S$ and $HgS.Na_2S$ (p. 500). From such solutions the mineral is easily formed as shown by Allen and Crenshaw. Cinnabar is not attacked by dilute solutions of sulphuric acid nor by a mixture of ferric and ferrous sulphate, but it is dissolved by dilute hydrochloric acid and is readily attacked by nascent chlorine. The rare selenides tiemannite (HgSe) and onofrite (Hg(SSe)) are probably primary.

It is apparent that the enrichment by oxidation of cinnabar deposits is not easily accomplished and the secondary minerals are usually confined to a little native metal, which is often found in drops on cinnabar and probably is reduced by the hydro-

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[1] *Trans.*, Am. Inst. Min. Eng., vol. 47, 1913, p. 217.

[2] E. T. Allen and J. L. Crenshaw, The sulphides of zinc, cadmium and mercury, *Am. Jour. Sci.*, 4th ser., vol, 34, 1912, pp. 367–383.

W. H. Emmons, Enrichment of ore deposits, *Bull.* 625, U. S. Geol. Survey, 1917, pp. 392–398.

F. F. Grout, *Econ. Geol.*, vol. 8, 1913, p. 427.

T. M. Broderick, Some experiments bearing on the secondary enrichment of mercury deposits, *Econ. Geol.*, vol. 11, 1916, pp. 645–651.

carbons which are common in many deposits; and occasionally to a little mercurous chloride or calomel. It is not impossible that some of the mercury may be of primary origin for G. F. Becker in some of his experiments[1] obtained precipitates of mixed sulphide and native metal. The oxidized products of cinnabar consist besides native metal and calomel (HgCl), of the red montroydite (HgO) and a number of oxychlorides, such as terlinguaite (Hg_2ClO) and eglestonite (Hg_4Cl_2O). They are generally found in the deposits occurring in arid, wind swept regions, where sodium chloride abounds. Large masses of montroydite, calomel and oxychlorides occurred at Terlingua,[2] Texas, but even here cinnabar is the principal mineral. According to Hill,[3] pyrolusite and pyrite are quite abundant in the deposits. All the conditions are thus present for the development of nascent chlorine so that the cinnabar can be transformed into $HgCl_2$ which again by reaction with the abundant calcite would produce oxychlorides.[4]

The mercurous sulphate is only slightly soluble in water (0.058 grams per liter), and the mercurous chloride HgCl is still less soluble (0.002 grams per liter) while $HgCl_2$ is easily soluble.[5] Though the experiments seem to indicate the insolubility of cinnabar in sulphuric acid it will be recalled that G. F. Becker found cinnabar, at New Idria, in a crust of epsomite and other soluble sulphates.[6]

It has been observed that when mercurial tetrahedrite occurs in a deposit, cinnabar may be found in the oxidized products. This is shown, for instance, in several mines near Sumpter, Oregon.[7] It is not known by what reactions this is accomplished.

Secondary Sulphides of Quicksilver.—From the position of quicksilver in the table of Schürmann's reactions it would appear that in the absence of oxygen the deposition of secondary sulphides should be easily effected. Broderick's experiments showed, however, that the sulphide was not precipitated from

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[1] *Am. Jour. Sci.*, 3d ser., vol. 33, 1887, p. 199.

[2] W. F. Hillebrand and W. T. Schaller, *Bull.* 405, U. S. Geol. Survey, 1909.

[3] B. F. Hill, The Terlingua quicksilver deposits, *Bull.* 15, Univ. Texas, 1902.

[4] T. M. Broderick, *op. cit.*

[5] W. H. Emmons, *op. cit.*, p. 393.

[6] *Mon.* 13, U. S. Geol. Survey, 1888, p. 307.

[7] W. Lindgren. *Twenty-second Ann. Rept.*, U. S. Geol. Survey, pt. 2, 1901, p. 708.

chloride solutions by marcasite, pyrite or realgar. A precipitate was obtained with stibnite and chalcopyrite. Cinnabar replacing sulphides have not, however, been observed in nature. There exists a black amorphous modification of HgS, known as metacinnabar, which has been found in the oxidized zone of many deposits. Allen and Crenshaw have ascertained that this is an unstable form and have obtained it artificially by precipitating dilute, acid solutions of mercuric salts by sodium thiosulphate, a reaction which might well take place in nature. Cinnabar was obtained only from alkaline solutions. At 100° C. the black sulphide changes into cinnabar.

Cadmium.—Cadmium is commonly contained in zinc blende, probably as an isomorphous mixture. The cadmium sulphate is as easily soluble as zinc sulphate and it is present in some mine waters of Joplin, Missouri, and Butte, Montana (p. 905). Cadmium sulphide, greenockite,[1] is often found as coatings on zinc blende in agreement with Schürmann's rule, and it occurs in this way even in the zone of oxidation.

Nickel and Cobalt.—Nickel contained as silicate in basic igneous rocks is concentrated during oxidation to hydrous nickel silicates like garnierite. Under similar conditions the cobalt separates as asbolite, a black earthy hydrous oxide with manganese. Both are colloid precipitates.

The primary nickel ores consist of niccolite (NiAs), pentlandite ((FeNi)S), polydymite (Ni_4S_5), chloanthite ($NiAs_2$) and minerals allied to the latter form. Millerite (NiS) is of no economic importance.

The nickel sulphides are easily attacked by oxygen, sulphuric acid and ferric sulphate. Pyrrhotite with which they usually occur is also easily decomposed.

Nickel sulphate ($NiSO_4$) is easily soluble and does not hydrolyze like iron sulphate to form the trivalent oxide. Therefore, nickel and iron are separated during oxidation. Much of the iron remains in the gossan while the nickel is carried away in solution.[2] Pyrrhotite ores are likely to lose nickel if exposed to oxidation in stopes and piles.

If arsenic is present stable arsenates may be formed in small quantities, in the oxidized zone; like the apple green annabergite

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[1] E. T. Allen and J. L. Crenshaw, The sulphides of zinc, cadmium and mercury, *Am. Jour. Sci.*, 4th ser., vol. 34, 1912, p. 341.

[2] W. H. Emmons, *Bull.* 625, U. S. Geol. Survey, 1917, p. 460.

($Ni_3As_2O_8+8H_2O$). In a similar way, erythrite ($Co_3As_2O_8+8H_2O$) develops from cobalt arsenides. Basic sulphates of nickel are not known. There is some evidence that millerite and polydymite may be locally precipitated as supergene sulphides. Both nickel and cobalt are often found in acid mine waters.

The more common cobalt arsenides and sulpharsenides include smaltite ($CoAs_2$), cobaltite ($CoAsS$) and linnaeite (Co_3S_4). These minerals are easily attacked by reagents and the sulphate being very soluble the metal is readily dissipated during oxidation. The hydrous oxide, asbolite, seems to be the only stable oxidized mineral, except the unimportant arsenate.

Chromium.—The primary chromite of igneous deposits is very insoluble and often remains in detrital deposits. Chromium mica (mariposite and fuchsite) is known in sulphide deposits where they traverse serpentine or peridotite. Chromates such as the red crocoite ($PbCrO_4$) may be found in the oxidized zone. Possibly chromite is slightly affected by ferric sulphate solution. Chromium sulphates of green or purple color have been found in quicksilver veins traversing serpentines.

Manganese.—The primary manganese minerals in ore deposits are manganosiderite ($(MnFe)CO_3$), more rarely rhodochrosite ($MnCO_3$) or rhodonite ($MnSiO_3$), still more rarely alabandite (MnS) or hauerite (MnS_2). Hausmannite (Mn_3O_4) occurs in some deposits of igneous metamorphism.

Manganese is more soluble than iron and less easily precipitated. Calcite which readily precipitates iron has little effect on manganese solutions. These facts explain the separation of iron and manganese in oxidized deposits.

The universal products of oxidation are pyrolusite (MnO_2) or its hydrated derivatives like psilomelane and wad. Manganite ($Mn_2O_3.H_2O$) and braunite ($3Mn_2O_3.MnSiO_3$) in places accompany pyrolusite. Many of the hydrated forms of manganese dioxide contain in chemical combination the oxides of lead, copper, zinc, cobalt, barium and potassium and are probably hardened colloid precipitates.

Tin.—In tin deposits cassiterite is ordinarily the most abundant ore mineral and exhibits great resistance to solution and transportation. Frequently it remains after other constituents have been dissolved, the outcrops appearing enriched in tin. Some of the surface croppings of the Freiberg veins are reported to have contained considerable amounts of cassiterite, probably, according to Stelzner, released from the zinc blende in which it was disseminated as minute crystals. Above the water level the Cornwall veins contained mainly tin, the accompanying copper having been leached. Doelter[1] has shown experimentally that cassiterite is perceptibly soluble in water, but this is denied by J. P. Goldsberry.[2] According to some authorities[3] on Cornwall, cassiterite is occasionally found as a cement in gravels and as impregnations in long-buried deer horns. Lately Scrivenor has shown that the cassiterite is simply carried mechanically into the cavities of the antlers. The occurrence of pebbles of wood tin—a fibrous form of cassiterite—is reported from placers of Saxony, Cornwall, several places in the western states (as at Wood's Creek, Montana), and Bolivia. Stelzner regarded the Bolivian pebbles as derived from stannite (Cu_2FeSnS_4) or from stanniferous pyrite. The cassiterite had apparently not been altered.

Tungsten.—Scheelite ($CaWO_4$) and wolframite $(Fe,Mn)WO_4$ are common in many deposits, especially in those formed at high temperatures. A sulphide of tungsten (WS) looking like molybdenite has recently been discovered in Utah.[4] These primary minerals are very resistant and often occur in placers. Tungsten minerals are slightly attacked by dilute sulphuric acid.[5] The solutions are easily hydrolyzed to insoluble oxides, which also are precipitated by ferric salts, acids, etc. Tungstite[6] ($WO_3.H_2O$) usually results and forms a canary yellow coating.

Vanadium.—Few primary vanadium minerals are known. Igneous rocks, according to F. W. Clarke, contain an average of 0.018 per cent. vanadium and many pyroxenes and micas carry notable amounts. Roscoelite is found in many gold quartz veins. It is also widely transported in anorganic and organic cycles and thus found in sediments, clays, coal, etc.

Vanadates, like vanadinite ($PbClPb_4(VO_4)_3$) or descloizite ($4\,(Pb,Zn)O.V_2O_5H_2O$), are present in the oxidized zone of many lead deposits of igneous affiliations and may occasionally be

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[1] *Min. pet. Mitt.*, vol. 11, 1890, p. 325.

[2] W. H. Emmons, *Bull.* 625, U. S. Geol. Survey, 1917, p. 399.

[3] J. H. Collins, *Min. Mag.*, vol. 4, 1880, pp. 1 and 103; vol. 5, 1883, p. 121.

[4] R. C. Wells and B. S. Butler, Tungstenite, a New Mineral, *Proc.*, Washington Acad. Sci., vol. 7, 1917, pp. 596–599.

[5] R. W. Gannett, *Econ. Geol.*, vol. 14, 1919, pp. 68–78.

[6] F. L. Hess, Tungsten minerals and deposits, *Bull.* 652, U. S. Geol. Survey, 1917, p. 34.

of economic importance for the vanadium contained, which probably is derived from the country rock.

Regarding the complex vanadium oxysalts in deposits in sandstone and the rare vanadium sulphide patronite see pp. 407–412. The former include vanadium sulphates and uranium vanadates, and vanadium mica (roscoelite). These are undoubtedly deposited by meteoric waters, but the history of derivation and succession is imperfectly known.

Molybdenum.—From molybdenite the common primary ore mineral, molybdite (MoO_3) or molybdic ocher[1] ($Fe_2O_3.3MoO_3.7\frac{1}{2}H_2O$) is formed as secondary yellow powder. The lustrous orange plates of wulfenite ($PbMoO_4$) are common in the oxidized zone of deposits containing galena and molybdenite, and is in places as at Mammoth, Arizona, of economic importance; rarely it is accompanied by powellite ($CaMoO_4$). W. H. Emmons states that molybdenite is not attacked by H_2SO_4 or HCl, and not even by ferric sulphate. The metal is not very mobile and is not commonly concentrated in the oxidized zone; neither does it appear in the supergene sulphide zone. Molybdenite does oxidize however. It appears also as the readily soluble blue ilsemannite which by later investigations[2] appears to be molybdenum sulphate. It appears in some mine waters to which it may give a deep blue color.

Most probably this mineral is of colloidal origin and has been formed from molybdite or molybdic ocher by colloidal solution methods. Attention is also directed to the presence of molybdenum in vanadium deposits in sandstone, which probably were formed by meteoric waters and to the reported existence of a colloidal sulphide.[3]

Bismuth.—Among the primary bismuth minerals, bismuthinite (Bi_2S_3) the native metal and various lead-bismuth sulphides are the most important. The principal oxidized minerals of bismuth are bismite, $Bi_2O_3.3H_2O$, bismutite, $Bi_2O_3.CO_2.H_2O$ and several arsenates, for instance, arsenobismite ($2Bi_2O_3.As_2O_5.2H_2O$). In general the bismuth salts are difficultly soluble in water and the metal is in this respect like lead. They do not show great mobility and bismuth is not found abundantly in the supergene sulphide zone. Nevertheless it seems certain that

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[1] W. T. Schaller, *Am. Jour. Sci.*, 4th ser., vol. 22, 1907, p. 297.

[2] W. T. Schaller, *Jour.* Washington Acad. Sci., vol. 7, July 19, 1917.

[3] F. Cornu, *Zeitschr. Chem. Indust. Kolloide,* vol. 4, p. 190, 1909.

bismuth has been concentrated in places in the zone of direct oxidation as shown by masses of bismutite and hydrous bismuth arsenate (arsenobismite) found at Tintic[1] and elsewhere. Other observations point the same way: At the Great Cobar[2] copper mine (p. 698), in New South Wales, the smelting records show bismuth to have been abundant in the zone of oxidation while there is very little of it in the primary ore.

Another evidence of transportation of the metal is its occasional presence with lead in basic iron sulphates like jarosite.

Arsenic.—The primary arsenical minerals in ore deposits are arsenopyrite (FeAsS), smaltite ($CoAs_2$), tennantite ($Cu_8As_2S_7$), enargite (Cu_3AsS_4), proustite (Ag_3AsS_3), and many rarer sulpharsenides of copper, silver and lead.

Supergene minerals of arsenic include native metal, arsenolite (As_2O_3), mimetite ($PbCl.Pb_4As_3O_{12}$), olivenite ($Cu_3As_2O_8.Cu(OH)_2$), realgar (AsS), orpiment (As_2S_3), scorodite ($FeAsO_4.2H_2O$), pharmacosiderite (basic, hydrous, ferric arsenate) proustite and pearceite (Ag_9AsS_6).

Arsenic acts as an acid forming compound similar to phosphorous. The reactions in acid solution are probably carried out by means of alkaline arsenates and arsenites. The sulphides, which are almost insoluble in dilute H_2SO_4, easily form soluble double salts with alkaline sulphides, *e.g.*, Na_3AsS_3. On the whole arsenic is fairly mobile in the oxidized zone, more so than antimony. Arsenopyrite oxidizes to ferrous sulphate and arsenic trioxide:

$$2FeAsS + 11O = 2FeSO_4 + As_2O_3.$$

Finally scorodite and pharmacosiderite are formed. Arsenopyrite is attacked by ferric sulphate and sulphuric acid. Smaltite oxidizes to arsenates:

$$3CoAs_2 + 14O = Co_3As_2O_8 + 2As_2O_3.$$

Enargite is very resistant to oxidation but yields arsenate, sulphate and sulphuric acid:

$$2Cu_3AsS_4 + 35O + 5H_2O = Cu_3As_2O_8 + 3CuSO_4 + 5H_2SO_4.$$

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[1] W. Lindgren, *Prof. Paper* 107, U. S. Geol. Survey, 1919.
A. H. Means, *Am. Jour. Sci.*, 4th ser., vol. 41, 1916, p. 125.

[2] E. C. Andrews, Report on the Cobar copper and gold field, *Min. Res.* 17, Geol. Survey N. S. W., 1913, p. 113.

If there is not sufficient oxygen present, some chalcocite or covellite may form.

Native arsenic in concentric shells is not uncommon. Large masses occurred at Washington Camp, Arizona. It is apparently always a product of oxidation and was deposited by colloidal solutions. Films of secondary proustite occurred on arsenic in some deep mines at Freiberg, Saxony, which shows that the metal is not confined to the uppermost part of the oxidized zone.

Realgar and orpiment are probably always supergene sulphides but they are not found in the secondary zones of copper deposits. They are rather more characteristic of the oxidized zone, and often are derived from arsenopyrite. The chemistry of their deposition is uncertain.

Sulpharsenides of silver are characteristic of the supergene zone of many silver deposits.

In limestone the basic arsenates of copper are easily fixed; calcium, iron and zinc also enter in their composition.

In non-calcareous rocks the arsenates are scarce and in fact, in many oxidized zones and supergene sulphide zones the arsenic of the primary enargite ore is almost wholly removed. This is the case at Chuquicamata, Chile, and at Butte, Montana.

Antimony.—The primary minerals of antimony in ore deposits are stibnite (Sb_2S_3) and a considerable number of sulphantimonides in part of copper like tetrahedrite ($4Cu_2S.Sb_2S_3$), and famatinite ($3Cu_2S.Sb_2S_5$); in part of lead like jamesonite ($Pb_2Sb_2S_5$), and many others; in part of silver like pyrargyrite (Ag_3SbS_3).

The supergene minerals comprise native metal, cervantite (Sb_2O_4), senarmontite (Sb_2O_3) and valentinite (Sb_2O_3); further, oxysalts like bindheimite ($Pb_3Sb_2O_8.H_2O$) and stibiconite ($2SbO_2.H_2O$); finally many silver sulphantimonides of which stephanite (Ag_5SbS_4) and polybasite (Ag_9SbS_6) are the most important. Antimony is often associated with gold, silver and lead.

Antimony is considerably less mobile than arsenic during oxidation and as the salts hydrolyze in water there is a strong tendency to form insoluble white or yellowish oxides. To these stibnite usually alters and once formed they are difficultly moved. When, as often happens, lead is present in antimony deposits insoluble compounds like bindheimite are likely to develop.

Native antimony is not common but is probably of supergene origin.

Stibnite is very slightly attacked by dilute sulphuric acid and by ferric sulphate, but reacts readily with alkaline solutions and like mercury and arsenic forms soluble alkaline sulpho salts.

Though according to Schürmann's series antimony sulphide would be expected to replace various other sulphides no examples of this have been found. Stibnite is, however, often found as hairlike crystals in vugs where the mineral certainly is of very late origin. The same applies even more to certain occurrences of capillary and feathery jamesonite but whether this is the effect of late hypogene or deep supergene solutions is not known definitely.

According to W. Malcolm[1] stibnite is being deposited at the present time in the West Gore mine, Nova Scotia, and also a "red sulphide" perhaps kermesite (Sb_2S_2O) is also said to be forming, both probably from alkaline water.

Antimony is thus not extensively transported in the supergene sulphide zone. It becomes of importance only in the case of silver sulphantimonides (p. 884).

MINE WATERS[2]

Chloride Waters

In ores free from sulphides and other easily decomposed minerals the mine waters differ little from the ordinary ground waters of the region. Examples of such waters are found in the iron and copper mines of Michigan. In the following table analyses 1, 2, and 3 represent average waters of the upper circulation; 4 and 5 give the composition of the deep waters of the same region, which differ very materially from the shallow waters and contain an abundance of calcium chloride. The deep waters in the copper region contain some copper, zinc, nickel, and traces of boron.

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[1] *Mem.* 20-E, Geol. Survey Canada, 1912, p. 296.

[2] W. H. Emmons and G. L. Harrington, A comparison of the waters of mines and of hot springs, *Econ. Geol.*, vol. 8, 1913, pp. 653–659.

E. T. Hodge, The composition of waters of sulphide ores, *Econ. Geol.*, vol. 10, 1915, pp. 123–139.

ANALYSIS OF MINE WATERS FROM UPPER AND LOWER LEVELS OF IRON AND COPPER MINES OF MICHIGAN

[Parts per million]

	1	2	3	4	5
Cl	3.5	18.68	6.00	25,360	176,027
Br	....	....	....	....	2,200
CO_3	24.2	163.00[1]	41.60	Not det.	....
SO_4	18.8	13.14	12.10	1,045	110
Ca	12.9	62.29	15.20	7,902	86,478
Mg	2.0	28.20	9.60	566	20
Na	11.3	19.00	4.00	7,290	15,188
K	....		1.50		411
Al_2O_3	4.0	18.20	0.60	700	10
Fe_2O_3			1.23		
Mn	....	....	....	....	4
Cu	....	....	....	....	16
Ni	....	....	....	....	6
SiO_2	14.5	9.80	8.43	....	20
Loss	2.8	....	....	....	....
Total	94.0	332.31	100.26	42,863	280,490

[1] CO_2.

1. Mass copper mine, Michigan. Water from upper levels. Analyst, Dearborn Chemical Works, Chicago. A. C. Lane, Mine waters. Thirteenth Annual Meeting, Lake Superior Mining Institute, June, 1908, p. 31.
2. Vulcan iron mine, Michigan. Analyst, G. Fernekes. *Idem*, p. 6.
3. Newport mine, Gogebic district, Michigan. Analyst, R. D. Hall. Free CO_2 18.0. Residue dried at 100° C. 108.30. Van Hise and Leith, *Mon.* 52, U. S. Geol. Survey, 1911, p. 543.
4. Republic iron mine. Seventeenth level. Temperature 57° F. Analyst, G. Fernekes. Calculates to NaCl 18,510 and $CaCl_2$ 21,800. A. C. Lane, *op. cit.*, p. 10.
5. Quincy copper mine. Drippings on fifty-fifth level. Similar waters from the Calumet & Hecla mine also contain some zinc. Analyst, G. Fernekes. Chiefly calcium and sodium chlorides and sodium bromide. *Op. cit.*, p. 48. Trace boron and strontium. No barium or carbon dioxide.

Occasionally mine waters comparatively rich in silica may be encountered. Thus, E. T. Allen analyzed a surface water from the Mesabi iron range, Minnesota, which contained 22 parts per million of silica and only about 14 parts of sulphates and 5 parts of carbonates of calcium and alkali metals.[1]

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[1] *Mon.* 43, U. S. Geol. Survey, 1903, p. 264.

Salt waters, containing mainly sodium chloride to the amount of several per cent., are reported from the Kalgoorlie mines in Western Australia, where they began to come in at water level, a few hundred feet below the surface. At the Great Boulder Proprietary the water contained 9 per cent. of sodium chloride and also much magnesium chloride.[1] Similar waters are now coming into the deep levels of the Bendigo mines.[2] Sulphides are not abundant in these mines. The analysis of a sample taken 4,280 feet below the surface, in the Victoria Reef quartz, where the temperature of the water is 114° F., is as follows, contained in parts per million:

NaCl, 1,308.45; Na_2SO_4, 75.79; Na_2CO_3, 37.18; $CaCO_3$, 124.41; $MgCO_3$, 45.76; SiO_2, 21.45; $(Al, Fe)_2O_3$, 2.86; total, 1,615.90.

According to T. A. Rickard[3] the mine water at Mammoth, Pinal County, Arizona, contains 86 parts per million of sodium chloride, and that from Stratton's Independence mine, at Cripple Creek, Colorado, 51 parts of the same salt.

Carbonate Waters

The mine waters from the Wardner lead mines, in the Coeur d'Alene district, Idaho, are rich in ferrous carbonate (from siderite in the ore) and deposit abundant limonite. A sample from the Reed level, Bunker Hill & Sullivan mine, showed 70 parts per million of total solids, chiefly bicarbonate and sulphate of calcium.[4]

A number of analyses of waters from the lead mines of southeastern Missouri are given by E. R. Buckley.[5] The waters come from the La Motte sandstone and Bonneterre dolomite, generally at depths of a few hundred feet. The total solids are at most 500 parts per million, of which 200 parts or more are calculated as calcium-magnesium carbonates. The sulphates, calculated as the magnesium salt, are at most 200 parts per million, while sodium chloride averages only 50 parts. Silica is low. All contain a little lead, at most 1 part per million, calculated as

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[1] T. A. Rickard, Formation of bonanzas, *Trans.*, Am. Inst. Min. Eng., vol. 31, 1901, pp. 198–220.

[2] W. J. Rickard, Deep mining at Bendigo, *Mining Mag.*, 1910, p. 281.

[3] *Trans.*, Am. Inst. Min. Eng., vol. 31, 1901, pp. 198–220.

[4] F. L. Ransome, *Prof. Paper* 62, U. S. Geol. Survey, 1908.

[5] Missouri Bur. Geology and Mines, vol. 9, pt. 1, 1909, p. 249.

lead sulphate, and generally a trace of zinc. They are weak waters mainly on account of the small amount of pyrite in the deposit.

Sulphate Waters

Oxidation of Pyrite.—Where pyrite is present in notable quantities its oxidation materially changes the composition of the waters. The sulphuric acid radicle increases rapidly and displaces the equilibrium so that the normal calcium carbonate waters are changed into those containing mainly calcium sulphate. When the free sulphuric acid increases still further the water becomes rich in the sulphates of aluminum (by the decomposition of sericite and other silicates) and iron, the latter present as both ferrous and ferric sulphate. Free hydrochloric acid is sometimes present. In waters above or at the water level these sulphates may be present in large quantities. Below the water level free acid is rarely found and the sulphate of aluminum is absent. The iron is present as ferrous sulphate and diminishes in quantity with increasing depth. The characteristic calcium sulphate waters persist for wide spaces around pyritic deposits and also reach considerable depths. Besides the sulphates mentioned, the mine waters of the oxidized zone contain almost all the metals occurring in the deposit. Zinc sulphate is especially abundant; copper sulphate is usually present, lead much more rarely; arsenic is common and antimony rare.

The waters of coal mines show plainly the result of the oxidation of the pyrite and marcasite occurring in the beds. Such waters are often rich in the sulphates of ammonium, calcium, iron, and aluminum, and even in free sulphuric acid. In the drainage from the mines the iron appears as ferrous sulphate, from which, by oxidation, ferric sulphate is formed. Coarsely crushed coal washed with distilled water is said to yield free sulphuric acid in the filtrate. Mine waters from coal mines occasionally contain zinc, copper, cobalt, and nickel. A water from the coal mine of the Dravo-Doyle Company, in Pennsylvania, showed according to analysis by the Pittsburgh testing laboratory of the Bureau of Mines:

Free H_2SO_4	117 parts per million
$Fe_2(SO_4)_3$	4,970 parts per million
$Al_2(SO_4)_3$	140 parts per million
$FeSO_4$	54 parts per million

More or less of the sulphate of calcium and magnesium are also usually present.

Examples.—A series of analyses of the Comstock waters, Nevada, by J. A. Reid well illustrates the occurrence of sulphate waters. No. 3 is a concentrated sulphate water from the oxidized zone; Nos. 1 and 2 are deeper hot waters, resulting from the reaction between an ascending sodium-carbonate water and sulphuric acid from the upper zones. The ores contain mainly gold and silver and are not rich in pyrite.

ANALYSES OF MINE WATERS FROM THE COMSTOCK LODE

[Parts per million]

	1	2	3
Cl	1.27	19.00	127.60
SO_4	380.38	474.00	209,100.00
CO_3	115.03	20.45	
K	8.39	53.40	
Na	57.13	132.00	535.00
Ca	148.10	100.10	1,286.00
Mg	154.03	5.88	6,590.00
Al		1.37	9,670.00
Mn			885.10
Cu			147.50
SiO_2	30.50	133.40	616.00
Fe		6.33	5,025.02
H			2,575.00
Total salinity	764.40	965.60	110,958.30

1. Water from the 600-foot level of the Savage mine. G. F. Becker, *Mon.* 3, U. S. Geol. Survey, 1882, p. 152.

2. Waters from the C. & C. shaft at the 2,250-foot level. John A. Reid, *Bull.* California Univ. Dept. Geology, vol. 4, 1905, pp. 177–199.

3. Vadose water from the Central tunnel. *Idem.*

Some of the mine waters of the Joplin zinc region, where the deposits contain, besides zinc blende and galena, some pyrite or marcasite, are extremely rich in zinc sulphate and contain also the sulphates of iron and aluminum. (See analysis No. 1 in table on p. 905.)

The water of the Rothschönberger tunnel, draining the mines at Freiberg, Saxony, is a good example of a dilute mine water which has traversed the old workings of veins carrying pyrite, galena, and zinc blende. (See analysis No. 2.)

The same principle is illustrated by the analyses of two waters from the mines at Butte, Montana. No. 3 is from a deep level, but rather far from the principal vein system; No. 4 is from the 1,200-foot level in one of the principal mines; it has acquired the habit of a water of the upper oxidized zone because the water level has been artificially lowered and the oxidation of the pyrite is progressing rapidly. No. 5 is a deep water from Cripple Creek.

ANALYSES OF MINE WATERS

	1	2	3	4	5
Cl	2.7	12.4	6.8	13.0	0.8
HCO_3	....	....	13.5	....	210.0
SO_4	6,153.2	124.8	406.5	2,672.0	1,088.0
SiO_2	107.6	18.0	23.0	47.7	64.0
AsO_4	....	....	Trace.	....	....
Ca	345.3	46.4	151.2	132.5	564.7
Mg	25.2	14.5	28.2	61.6	22.4
Na	49.9	....	16.2	39.6	51.2
K	0.5	....	7.1	13.1	7.1
Fe''	474.6	6.6	1.8	....	....
Fe'''			....	159.8	....
Mn	1.7	....	0.5	12.0	....
Zn	2,412.0	8.9	0.3	852.0	....
Cd	9.0	....	....	41.1	....
Al	142.1	....	....	83.5	....
Cu	3.7	....	Trace.	59.1	....
Co + Ni	....	....	....	0.5	....
Sn	....	....	....	17.0	....
	9,727.5	231.6	655.1	4,204.5	2,308.8

1. Water from Alabama Coon mine, Joplin, Missouri. H. N. Stokes, analyst.

2. Water from the Rothschönberger Stolln, Freiberg, Saxony, at point of discharge. Analysis by Frenzel. Recalculated by F. W. Clarke, Data of geochemistry, *Bull.* 616, U. S. Geol. Survey, 1916, p. 632.

3. Water from 2,200-foot level, Green Mountain mine, Butte, Montana, remote from veins. W. F. Hillebrand, analyst.

4. Water from 1,200-foot level, crosscut St. Lawrence, Butte, Montana. Tin possibly accidentally introduced? Faintly acid. W. F. Hillebrand, analyst. Fe'' probably changed to Fe''' during exposure to air.

5. Cripple Creek, Colorado, water from El Paso Tunnel draining the lowest workings. R. C. Wells, analyst. Fe + Al, etc., 0.6. Free CO_2 trace.

F. L. Ransome[1] mentions a mine water from Goldfield, Nevada, which contained about 4,250 parts per million of total solids, mostly sulphates of iron, sodium, magnesium, and calcium. The silica in such waters is generally low.

A. C. Lawson[2] describes the mine water from the Ruth mine 335 feet below the surface, in the chalcocite blanket in the porphyry of Ely, Nevada. The temperature was 16° C., decidedly higher than the average annual temperature of the region. The total solids were 1,094 parts per million, of which 359 parts were calculated as calcium sulphate, 130 as magnesium sulphate, 93 as alkaline chlorides, 160 as ferrous sulphate, and 7 as ferric sulphate.

Two analyses of the mine waters at the copper mines of Cananea, Mexico, have been received through the courtesy of Mr. W. H. Emmons. The waters come from an upper and a deeper level and have percolated through a sericitized rock with a considerable amount of chalcocite and pyrite, though there are no solid masses of pyrite.

ANALYSES OF WATERS FROM THE CAPOTE MINE, CANANEA, MEXICO

F. G. Hawley, Analyst

[Parts per million]

	300-foot level.	900-foot level.
SO_3	4,220	3,714
Cl	Not determined.	22
H_2SO_4 (free)	970	nil.
SiO_2	76	56
FeO	393	674
Al_2O_3	Not determined.	42
CaO	610	1,053
MgO	102	144
MnO	305	198
ZnO	Not determined.	315
CuO	2,097	76
$(K,Na)_2O$	Not determined.	198

Fe almost wholly as Fe″.

H_2SO_4 not subtracted from total SO_3.

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[1] Geology and ore deposits of Goldfield, Nevada, *Prof. Paper* 66, U. S. Geol. Survey, 1909, p. 258.

[2] *Bull.* California Univ. Dept. Geol., vol. 4, 1906, pp. 287–357.

The deeper waters contain much more calcium sulphate as well as ferrous sulphate, but much less copper. Chalcocite probably reduces the ferric sulphate to ferrous.

Another instructive series was collected by W. H. Emmons and Laney at Ducktown, Tennessee. Here the water level is high and the ores consist of heavy masses of pyrrhotite with some chalcopyrite.

ANALYSES OF MINE WATER FROM DUCKTOWN, TENNESSEE[1]

R. C. Wells, Analyst

[Parts per million]

	1	2	3
SO_4	6,664.0	415.8	474.8
Cl	0.1	0.7	0.4
SiO_2	55.6	37.0	49.9
H_2SO_4 (free)	129.6	210.2	97.5
Al	433.0	14.5	19.1
Fe″	2,178.0	71.4	89.2
Fe‴	nil.	20.3	55.9
Mn	0.2	0.2	0.1
Ca	67.6	19.7	30.4
Mg	40.6	5.2	6.2
Cu	312.1	28.1	11.0
Zn	199.8	2.4	2.9
K	19.8	2.7	2.2
Na	23.4	5.2	5.5

1. Burra-Burra mine. Circulating water dripping from roof of drift just below chalcocite zone.
2. Callaway shaft, standing water, at water level, 90 feet below surface. Both 1 and 2 were collected with special precautions and sealed to prevent oxidation.
3. Callaway shaft, standing water 37 feet below water level.

All three waters are rich in free acid, the deepest sample being the least acid. The water of No. 1, collected just below the chalcocite zone, is extremely rich in sulphates but contains no ferric sulphate, while the water standing in the shaft contains both ferrous and ferric sulphate. There is less copper in the lower part of the standing body of water than at the surface but more calcium.

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[1] Quoted in *Bull.* 529, U. S. Geol. Survey, 1913, pp. 60–61.

The Homestake mine, South Dakota[1] is working a large lenticular body of gold ore with some pyrite, pyrrhotite, quartz, and amphibole. The ordinary creek waters are of the normal calcium carbonate type, and a salinity of about 300 parts per million. The ordinary mine water from the upper levels has a salinity of 510 parts and contains $CaSO_4$ and $(Ca,Mg)CO_3$. The deeper waters, from the 1,100 and 1,550 foot levels, have a higher salinity, as much as 1,228 parts, caused mainly by increase of $CaSO_4$. The salinity increases during dry periods.

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[1] W. J. Sharwood, *Econ. Geol.*, vol. 6, 1911, pp. 738–744.

CHAPTER XXXII

METALLOGENETIC EPOCHS[1]

INTRODUCTION

Wherever universal geological processes are in operation—weathering, sedimentation, metamorphism, deposition by underground waters and vulcanism—there mineral deposits may be forming as they have done during the long ages of the earth's history. Over larger or smaller areas the conditions may at a given time be favorable for the deposition of useful minerals. Such areas are called minerogenetic or metallogenetic provinces. The Jurassic iron ores of England and France, the bauxite deposits of the southern Appalachian States, the salt deposits of Kansas, the gold-quartz veins of California occupy minerogenetic provinces. The provinces may widen into minerogenetic regions: Thus the Cordilleras of the Americas form such a region marked by gold and silver deposits of igneous genetic affiliations. In the same way the Paleozoic of the whole Mississippi Valley region is a metallogenetic region characterized by lead and zinc deposits related in origin to circulating meteoric waters.

The time intervals favorable for the deposition of certain useful substances are called minerogenetic or metallogenetic epochs. Usually speaking they are short and transitory but they likewise may widen into periods and eras. Weathering and sedimentation may go on for periods but the mineral deposits due to these processes commonly represent a comparatively brief time. The Clinton hematites, for instance, though widely spread, were

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[1] This conception was first developed by L. De Launay who has contributed much to the study of this subject.

L. De Launay, Gîtes metallifères, vol. 1, 1913, pp. 241–288.

W. Lindgren, Metallogenetic epochs, Econ. Geol., vol. 4, 1909, pp. 409–420.

W. Lindgren, Gold and silver deposits in North and South America, *Trans.*, Am. Inst. Min. Eng., vol. 55, 1917, pp. 883–909.

formed during a limited period of littoral sedimentation and are covered and underlain by heavy Carboniferous and Ordovician limestones.

This applies even more to deposits of igneous affiliations which often were formed during a very brief interval when the emanations of each igneous phase found opportunity for escape and transportation. Each igneous province usually includes many smaller elements in which successive intrusions and effusions were followed by short epochs of various metallizations.

Each mineral deposit has usually a well marked paragenetic history due to the cooling or changing of solution. In any older deposit subsequent epochs of mineralization may have left their imprints, or geologic processes like metamorphism and weathering may have modified them.

The minerogenetic development of the several continents is outlined in briefest form in the following paragraphs. It is a fascinating subject for it connects in logical order the science of general geology with that of mineral deposits.

Main Epochs.—In the nature of the case metallogenetic epochs of weathering, sedimentation and erosion are not confined to any particular geological period. On the other hand the deposits genetically connected with igneous rocks and metamorphism are most abundantly formed during the great periods of such disturbances which are more or less closely affiliated with folding and mountain building. In his studies on European deposits, De Launay divides them into those of: 1. Pre-Cambrian age; 2. Hercynian age; and 3. Tertiary age.

It is necessary to realize, however, that the pre-Cambrian deposits, though generally formed at high temperature and great depth, include many ages far apart. The Hercynian deposits were formed during the great orogenetic disturbance, which falls between the end of the Paleozoic and the beginning of the Trias and which was accompanied by many intrusions. To this period belong such deposits as the tin veins of Cornwall and the silver-lead veins of Freiberg. "Hercynian" is a term that has been applied rather loosely and even metallogenetic epochs of distinctly Paleozoic age or of Triassic have been included in it. The Tertiary deposits are those which were formed during or following the great Alpine foldings and subsequent igneous activity. Folding is, however, a process not necessarily connected with igneous activity, and, therefore, we shall place less emphasis than De Launay has done on the connection between such tangential thrusts and mineral deposition.

EUROPE

Pre-Cambrian Epochs.—We find the pre-Cambrian deposits mainly in the great shield of Fenno-Scandia, in Sweden, Norway and Finland. They carry mainly iron and copper, more rarely lead and zinc or tin. Gold and silver are very scarce. Deposits of contact-metamorphic or magmatic type prevail with structural and mineralogical features indicating depth and high temperature. Dynamic metamorphism frequently has been superimposed. Iron deposits of sedimentary origin in part a little similar to those of North America are found in places.

Paleozoic Epochs.—Sedimentary ores of hematite and iron silicates are rather widespread in Europe. We find them in the metamorphosed Cambro-Silurian of northern Norway, in Great Britain, France, Germany, Bohemia and Spain. Among the deposits of igneous origin we note the copper and nickel ores of northern Norway, connected with gabbros intrusive in the Cambro-Silurian complex.

Hercynian Epochs.—The Hercynian movements fell between the Carboniferous and the Trias and resulted in great mountain chains extending East and West across Europe. Intrusions of granitic rocks of that age are laid bare by erosion which has also truncated the deposits so that the veins which form the dominant type present the high temperature or intermediate types. Tin, lead, zinc, copper and silver prevail among the metals; antimony and arsenic begin to appear. Iron deposits are also found but do not attain the importance of those of Fenno-Scandia. In this age were formed the tin deposits of England, France, Saxony, and Spain all connected with persilicic intrusives.

The majority of the deposits of the Iberian Peninsula (including those of Rio Tinto and the Mesa Central[1]), of the Central Plateau of France and of Central Germany, belong to this age.

The Ural Mountains are also regarded as a Hercynian range truncated by erosion and at least many of its deposits of iron, copper and gold are considered as induced by Hercynian intrusives. The deposits are generally of a deep-seated type.

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[1] The quicksilver deposit of Almaden should probably be included among these.

Permo-Triassic Epochs.—Between the Hercynian ranges extended desert plains and saline lakes or bays. The salt deposits of central Germany, the sedimentary copper ores of the Mansfield basin, the copper ores disseminated in sandstone in Germany and Russia belong to these epochs.

Jurassic and Cretaceous Epochs.—Broad transgressions of the seas with widespread formation of oölitic iron ores mark the Jurassic in England and on the continent. Glauconitic sands and chalk accumulated in the Cretaceous seas. The igneous forces were quiescent. Perhaps we do not err in attributing to the Jurassic the concentrations of lead-zinc ores in the Paleozoic of Belgium and in the Trias of Silesia.

Tertiary Epochs.—Few mineral deposits of northern Europe date from the Tertiary but southern Europe was the scene of great activity mainly in connection with the igneous outbursts which culminated south of the great Alpine arches and overthrusts.

1. In the areas occupied by the Mediterranean seas, we find widespread sulphur deposits and phosphates, both in the sedimentary beds.

2. The great Alpine intrusions of granitic rocks originated replacement deposits of siderite and magnesite in the Triassic and Paleozoic limestones together with many minor metal deposits. Among the latter should probably be placed the widespread lead-zinc deposits of the Alpine Trias.

3. Farther south, in Italy, we note the contact-metamorphic specularite deposits of Elba in connection with intrusive granite. The magnetite deposits of the Banat, likewise of contact-metamorphic origin also belong in this class.

4. The Eocene intrusion of greenstones (peridotites, gabbro, etc.) in the Alpine region and throughout the Mediterranean resulted in magmatic copper deposits, zeolitic copper deposits (Monte Catini), magmatic chromite deposits (Greece and Asia Minor) and contact-metamorphic emery deposits.

5. In Hungary and Transylvania, inside of the Carpathian arches, intrusions and effusions of andesite and dacite took place in the Miocene. Accompanying those was a metallization of gold and silver very similar to the Cordilleran types. These veins are of the type formed near the surface.

6. The last metallization of cinnabar accompanied the late Tertiary (and perhaps the present) igneous outbursts in Tuscany,

Austria, and Serbia. It is throughout of the type formed near the surface.

Some effects of the Tertiary metallization is noted in connection with eruptives in Bohemia, Saxony and the Hartz as well as in the Central Plateau of France. Lead, silver, nickel, cobalt, antimony are the principal metals supplied in these regions.

ASIA[1]

The mineral deposits of Asia have been described by De Launay, who in his work cited below, has collected a vast mass of data in the correlation of which there are still many blanks.

We find the pre-Cambrian deposits represented by gold-bearing veins in the Jenisei and perhaps the Lena districts in Siberia, and in Korea. Again we meet them in the great Indian platform, formerly connected with Africa. Here are deposits of gold-bearing quartz (Mysore), of iron, of manganese and of graphite.

Deposits of Hercynian or late Paleozoic age are found in many ranges in central Asia, such as the Ural, the Altai and the trans-Baikalian mountains. They are mostly well-defined veins carrying gold, lead, silver and other metals. To the same age are referred the tin and wolfram deposits of Malaya, Burma and Annam, many of which are accompanied by much later placers. Copper deposits in sandstone occur in northern Turkestan.

The deposits post-dating the Himalayan and Alpine systems are apparently rare. A few of them occur in the Caucasus. The Tertiary mineralization accompanying the igneous outbursts bordering the Pacific is, however, well represented in eastern Asia. We find gold-bearing veins of the Cordilleran type in Sumatra, Celebes, Borneo, the Philippines and Japan. In the latter country there is also a more varied mineralization of the same date including veins of copper, antimony and quicksilver.

AFRICA

In the main Africa consists of a vast platform of ancient rocks. Arabia, India and Madagascar have the same structure and were no doubt once connected with the African continent. The crystalline rocks are overlain by sediments, folded in places,

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[1] L. De Launay, Richesses minérales de l'Asie, Paris, 1911, pp. 816.

of pre-Cambrian and Paleozoic age; in the south these older rocks are covered by Karroo beds (Permo-Triassic). In the north, in Algeria and Tunis, there are Tertiary folds of the Alpine system with accompanying igneous rocks. Here we find a weak metallization with deposits of lead, zinc, antimony and quicksilver.

Tertiary and recent eruptives are absent from the main part of Africa except along a zone from Abyssinia to the Great Rift valley where volcanoes rise above the old plateau.

The old pre-Cambrian land mass contains abundant gold deposits, usually of the high temperature type as in Madagascar, in Rhodesia and in Swaziland, and gold is likewise found in wide distribution in the old conglomerates of pre-Cambrian age overlying the Swaziland crystalline rocks. Most famous among these conglomerates are those of the Witwatersrand, Transvaal, which furnish the larger part of the world's gold production. Somewhat later are perhaps the bed veins of Pilgrims Rest and other places in the Transvaal.

Other deposits of deep-seated origin in Rhodesia and Namaqualand carry chromite and copper ores of magmatic origin.

Tin deposits of pegmatitic or deep-seated vein type, with accompanying placers are found in the Transvaal and in Nigeria.

The copper deposits at Katanga, Belgium Congo, are economically important and are found in sandstone, shale and limestone of probably Paleozoic age. Their origin and time of formation are in doubt.

The latest mineralization in South Africa appears to be represented by the diamond-bearing peridotite pipes which are intrusive in Karroo series. They are probably of Cretaceous age.

AUSTRALASIA

On the old land mass of Australia no ore deposits (except gold placers) appear to have been formed since the beginning of Mesozoic time.

The pre-Cambrian era is marked by a wide-spread metallization of gold in western Australia; this is apparently associated with intrusive granites. In New South Wales we find the exceptional high temperature lead-zinc deposit of Broken Hill, which probably also dates from pre-Cambrian times.

In Victoria and adjacent parts of New South Wales, a rich metallization of gold, in quartz veins, marked the close of the

Silurian and also followed the intrusion of granitic rocks. The copper-gold ores of the Great Cobar,[1] in western New South Wales are probably also of early Paleozoic age but the deposits are isolated among sediments far from outcropping igneous rocks. Gold quartz veins of a similar age appear in the South Island of New Zealand.

South and West of the New England region,[2] in New South Wales, numerous intrusions accompanied by vein formation took place in the Devonian, while in New England and Queensland deposits of tin, tungsten and bismuth were formed during and after Permo-Carboniferous granitic irruptions.

Rich deposits of gold and silver of the Cordilleran type, also deposits of quicksilver, are found on the North island of New Zealand. They occur in the lavas of Tertiary age which form a part of the igneous girdle of the Pacific.

SOUTH AMERICA

In South America, as in the northern part of the continent, we must differentiate the eastern part in which strong mountain building forces have rested since the close of the Paleozoic era from the western margin, which is marked by the great Cordilleran Ranges, in which, since early Mesozoic times, the scenes have been set for a tremendous display of vulcanism and orogenic movements.

In the eastern part iron and gold are the principal metals. The former occurs in Brazil as sedimentary deposits of hematite of great extent and probably "Algonkian" age.

A little later than these and connected with intrusive pegmatites are the gold-bearing quartz veins of Brazil, mainly in Minas Geraes, and in the Guianas; all of these are of the high temperature type.

Along the Pacific Coast the pre-Cambrian is largely lacking and practically all of the deposits are of Tertiary age. A few of them, in Colombia and Chile, are of the type formed near the surface and carry gold and silver. The great majority are of the intermediate type and have suffered considerable erosion. Genetically they stand in connection with smaller masses of intru-

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[1] E. C. Andrews, Report on the Great Cobar copper and gold field, *Min. Res.* 17, Geol. Survey New South Wales, 1913, p. 188.

[2] *Idem.*

sive granite or diorite, as well as with rhyolitic and andesitic porphyries. The principal metals are copper and silver. Great lead and zinc deposits are not common. Gold is widely spread but even the greatest of the gold districts, in Colombia, do not compare with those of California and Australia.

Tin, with silver, occurs in a belt in the Cordillera Real in Bolivia. Tungsten deposits occur in Peru and Bolivia, quicksilver deposits in Peru.

Among the many districts which have yielded great treasures of silver we note particularly Cerro de Pasco in Peru, and Potosi, in Bolivia; at both places supergene enrichment has undoubtedly played an important part. A widespread mineralization of copper often accompanied by tourmaline is found in the central and northern parts of Chile.

A later epoch of mineralization verging upon the Recent has yielded superficial deposits of borax and nitrates probably derived from volcanic exhalations. Gold placers of Quaternary age have been worked at many places.

CENTRAL AMERICA

In Honduras, Nicaragua, San Salvador and Costa Rica—a region in which volcanoes and effusive rocks abound—we find many representatives of gold-silver veins of Tertiary age which were formed near the surface.

THE ANTILLES

A feeble mineralization of copper and gold is observed in the greater Antilles and in most cases it accompanied Cretaceous or Eocene intrusions of basic rocks.

NORTH AMERICA

In describing the metallogenetic epochs of North America it will be convenient to separate the eastern and western parts of the continent for, with the exception of their earliest history, they have little in common. The western or Cordilleran part including also almost the whole of Mexico, and Alaska contains few deposits older than the Cretaceous. The eastern part contains few that are more recent than that age and most of them are of pre-Cambrian age.

In the eastern part we distinguish the following metallogenetic epochs:

1. The Pre-Cambrian Epochs.—The pre-Cambrian era embraces a long time and many epochs of ore formation. Ages of deep weathering and sedimentation alternated with times of violent igneous action. The deposits of igneous affiliations are throughout of the deep-seated, high temperature types. The metals contained are iron, copper, nickel, gold and silver. Lead, zinc, antimony and quicksilver are rare. Naturally the deposits are best exposed in the great ice polished Canadian shield and adjacent parts of the United States, but many are also found along the Appalachian ranges, where the pre-Cambrian may be exposed.

Most ancient among the pre-Cambrian deposits are the "iron formations" contained in the sedimentary rocks associated with the Keewatin greenstones. At Vermilion, Minnesota, and many other places, hematite ores have developed by the oxidation of these iron formations.

But little more recent are the many magmatic and contact-metamorphic deposits connected with granitic intrusions in the Grenville series of sedimentary rocks in the Adirondacks and in Ontario. Among them we find titanic iron ores, magnetite, corundum, molybdenite, apatite, phlogopite. The iron ores occur in large deposits. The great zinc and manganese deposits of Franklin Furnace, New Jersey, should perhaps be correlated with these concentrations.

Iron formations were repeatedly laid down during the Huronian sedimentations in the Lake Superior region, but epochs of intrusions of granite and porphyry intervened and in connection with one of these—the Algoman epoch[1]—appears to have originated the majority of the gold-quartz veins of Ontario and Quebec. Though they are found in many districts they have been most successfully worked in the Porcupine region of central Ontario. Arsenopyrite is frequently associated with the gold.

The Pre-Cambrian epochs close with the great Keweenawan intrusions and effusions of basic rocks like gabbro, norite, diabase and basalt. These magmas were rich in copper, silver, nickel, cobalt and arsenic and many deposits of these metals were formed during this epoch. They are distributed over a wide area in the

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[1] W. G. Miller and C. W. Knight, Metallogenetic epochs in the pre-Cambrian of Ontario, *Twenty-fourth Ann. Rept.*, Ontario Bur. Mines, 1915, pp. 243–248.

Lake Superior region[1] and include the zeolitic type of native copper deposits, the silver-cobalt veins of Cobalt and the magmatic nickel-copper deposits of Sudbury.

2. **Paleozoic Sedimentary Epochs.**—During the long period of Paleozoic sedimentation in the Appalachian region at least three epochs were characterized by deposition of iron ores. We find oölitic hematite of Ordovician age in Newfoundland, and similar oölites in the widespread Clinton formation in the Silurian, the latter reaching their maximum development in Alabama. Less important siderites were deposited as "black bands" during the Carboniferous. The phosphate beds of Tennessee also belong here.

3. **Paleozoic Intrusives.**—From Nova Scotia to Alabama granitic intrusions took place during the early Paleozoic and some of them even extended into the Carboniferous. A mineralization of gold quartz veins followed these. We find these veins in Quebec, Nova Scotia, at various places in New England and particularly in the southern States from Maryland to Alabama. The age of the gold deposits in the latter region is still in dispute. Probably the majority of them are of Cambrian age. Many deposits of pyrite and chalcopyrite in this region appear to date from the same period.

4. **Paleozoic Epochs of Saline Deposits.**—Epochs of arid climates with accompanying development of deposits of evaporation recurred at several times during the Paleozoic era. Thus there are deposits of gypsum and salt in the Silurian of New York, in the Carboniferous of Michigan and in the Permian of Kansas and Texas.

5. **Epochs of Triassic Copper Deposits.**—The important period of the history of the igneo-genetic ores in the eastern part of the continent closed in the late Paleozoic. A feeble recurrence of ore formation took place during the early Mesozoic when the Triassic traps of New Jersey, Connecticut and the Bay of Fundy were injected as sheets or overflowed as lava streams. Copper ores of the zeolitic type and some contact-metamorphic iron and copper ores developed in places.

6. **Cretaceous and Later Periods of Lead and Zinc Concentration.**—Since the Triassic, vulcanism has rested and in the eastern part of the continent metal deposits have formed only by

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[1] C. R. Van Hise and C. K. Leith, The geology of the Lake Superior region, *Mon.* 52, U. S. Geol. Survey, 1911, p. 591.

the concentrating power of flowing surface waters, or of ground water in decaying rocks, or of ascending waters of meteoric origin.

In the Mississippi basin and in the Appalachian valleys the Paleozoic beds have been searched by saline and carbonate waters containing carbon dioxide and hydrogen sulphide. This solution and deposition has resulted in the many lead and zinc deposits which we find in the limestones of that region.

7. **Tertiary and Recent Periods of Rock Decay.**—For a long time, in fact since the end of the Paleozoic, large parts of this continent have been a land area and its rocks have been exposed to weathering and decay. Such periods have indeed not been absent during any extended time intervals and in the iron ores of the Lake Superior region as well as in certain manganese ores of Arkansas we find indications of long pre-Cambrian and shorter Paleozoic epochs of weathering and oxidation.

The mild climate which prevailed in the southern Appalachian region during the Tertiary favored the development of deposits caused by such agencies. The ores contain limonite, manganese dioxide, phosphates and bauxite.

In the Western or Cordilleran part of the Continent we distinguish the following metallogenetic epochs:

1. **The Pre-Cambrian Epochs.**—The western part of the Cordilleran region is lacking in great exposures of the pre-Cambrian but wherever they are present we find deposits of gold, iron and copper, rarely other metals. The correlation of these with the eastern epochs is often difficult but on the whole they are connected with late pre-Cambrian (Algonkian) intrusions of granite and subsilicic rocks. As may be expected, the deposits are of the high temperature type and consist mainly of veins and lodes. Among them may be mentioned the gold lodes of the Black Hills of South Dakota, the lenticular gold-quartz veins and copper deposits of Wyoming, many smaller deposits in New Mexico,[1] the gold, iron and copper deposits of certain parts of southern California and Arizona, near Yuma and Parker, copper deposits of the Jerome district in Arizona and many others. As in the East, lead, zinc and antimony are rarely found in these ores.

2. **The Early Mesozoic Epoch.**—Throughout the Paleozoic and the larger part of the Mesozoic the great interior province, now

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[1] W. Lindgren, L. C. Graton and C. H. Gordon, The ore deposits of New Mexico, *Prof. Paper* 68, 1910, pp. 48–51.

occupied by the Rocky Mountains, was the scene of almost uninterrupted sedimentation. The phosphate beds of Idaho date from this period. Not so along the Pacific Coast; for here we find in places evidence of intrusions and lava flows dating back to the early Paleozoic, and during the Trias and early Jurassic great effusions of subsilicic lavas took place. Copper deposits are found in these igneous rocks and many of these were doubtless formed shortly after this igneous activity. Among them we may probably count many minor deposits in California, the Bonanza deposit in the Copper River district, Alaska, and the ores of the Prince William's Sound in the same territory. In the Jurassic lavas of Vancouver Island, C. H. Clapp has found indication of a mineralization of the type formed near the surface.[1]

3. **The Late Mesozoic Epochs.**—The third and most important ore forming period followed and accompanied the great batholithic intrusions of the Pacific Coast to which an age from late Jurassic to early Cretaceous is generally assigned.

These intrusions of intermediate quartz-monzonitic or granodioritic character took place upon an immense scale and extended from Alaska to Baja California. The large batholiths of the Sierra Nevada and British Columbia are the most striking features but innumerable intrusions of smaller volume broke through the crust along both sides of the great batholiths and throughout Oregon, Idaho, southern California, southern Arizona, western Nevada, Washington and Alaska. One of the largest of these is exposed in the great granite area of central Idaho. But throughout this revolution and during the birth of mountain ranges on the coast the Cretaceous was being quietly deposited at sea level all over the eastern Cordilleran region from Canada to southern Mexico.

An epoch of intense metallization followed these intrusions, within the areas indicated. The great interior masses of the batholiths are usually free from deposits, as shown in the High Sierra, in the Clearwater region, and in British Columbia. But along their margins mineral deposits formed in abundance, as along the gold belt of California, and along the two contacts of the batholith of the Canadian and Alaskan coast regions. The latest researches by the Alaskan division of the United States Geological Survey indicate that the great placer fields of Alaska derived their gold from deposits of this epoch. Gold, primarily,

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[1] Econ. Geol., vol. 10, 1915, pp. 70–88.

and copper, secondarily, are the characteristic metals. Along the Pacific coast, where there is little limestone in the intruded sediments, lead is practically absent, but in the interior, as in Nevada and Arizona, where the intrusions came into contact with Paleozoic limestone, this metal, with zinc, begins to appear. Silver is everywhere present, but scarcely ever important, except where lead appears. Arsenic and antimony are not abundant, mercury is nowhere present in commercial quantities.

4. **The Early Tertiary Epoch.**—As if exhausted, the igneous forces appear to have rested until the close of the Cretaceous and then broke out in a new field, along the eastern margin of the Cordilleran region, at that time largely covered by a plastic mantle of Cretaceous shales and sandstones, several thousand feet in thickness, which rested on great accumulations of Paleozoic limestones.

The predominating magmas were again of intermediate character, and solidified as granular or porphyritic rocks, standing between the granites and the diorites; they contrast markedly with the potassic and acidic magmas of pre-Cambrian times. While it is not necessary to limit strictly this igneous activity to a certain time, there is little doubt that most of it took place in the Eocene. The eruptions mainly took the form of intrusions and largely that of laccoliths, undoubtedly because, unlike the conditions of the shattered rocks of the Pacific Coast, they were covered by this heavy, tough and still yielding mass of Cretaceous sediments. We find an enormous number of these intrusions at various horizons between the Cambrian and the Cretaceous or as dikes or stocks that break through the underlying pre-Cambrian. They are not comparable in extent to the great batholiths of the coast. They extend from British Columbia, through Montana, Utah, Nevada, Colorado, New Mexico, eastern Arizona, and attain their greatest development in eastern Mexico. For reasons already indicated many, perhaps most, of these intrusives never reached the surface. Only in a few cases, as in Montana and in Colorado, near Denver, do the strata of Laramie or Eocene age contain volcanic detritus.

The fourth Cordilleran epoch of metallization followed these intrusions; contact-metamorphic deposits and veins were formed in abundance around their margins. The characteristic metals are silver and gold with much lead and zinc, especially where the intrusions cut the limestones. Copper and iron are also present at such limestone contacts. Arsenic and antimony are far more in evidence than during the second epoch, but mercury is still absent. Rossland, Butte, Bingham, Tintic, and Leadville are representative districts.

5. **The Late Tertiary Epoch.**—Orogenic disturbances followed the intrusions; the whole Cordilleran region was lifted high above sea level, warped, and faulted. These disturbances may have facilitated sub-aërial eruptions; at any rate it is certain that the middle and close of the Tertiary witnessed outflows of lavas upon a magnificent scale.

These eruptions spread over large areas of the western part of the continent; less pronounced in British Columbia and Alaska, they are abundantly represented in California, Washington, Idaho, Colorado, Utah, Nevada, New Mexico, Arizona, and attained their greatest development in Mexico. Andesites and rhyolites are the predominating rocks. In some places the flows attained such thickness that during the later part of the volcanic epoch intrusions of magmas consolidated in them with granular structure.

During these eruptions, not strictly contemporaneous throughout, a fifth metallization took place, of which the characteristic metals are gold and silver. These deposits were often of great richness which is further accentuated by secondary processes; in fact most of the "bonanzas" belong to this class. Lead and zinc are not conspicuous except where the metallization took place in limestone. Copper is not abundant. Tellurium and antimony are common. Not that they are absent in older metallizations, but they, especially tellurium, seem particularly characteristic of this epoch. Quicksilver occurs in two belts, in Nevada and Mexico. The Comstock, Tonopah, Cripple Creek, Pachuca and El Oro are representative districts.

Large areas of volcanic rocks are barren. The metal deposits seem to have formed only near or at the foci of igneous activity, where connection could be established with underlying magmas. The most recent eruptions were mainly basalts, and these, except in one case, do not seem to have been affected by metallization.

6. **The Post-Pliocene Epoch.**—The youngest metallogenetic province is that of the Pacific Coast line. It is of very late age—Post-Pliocene apparently, and is characterized by mercury accompanied by few other metals. It developed in the Coast Ranges of California, following basalt eruptions and contem-

poraneous with it was a great development of hot springs. In part the deposition goes on at the present time.

Note that the quicksilver did not develop simultaneously with the birth of the Coast Ranges; these are much older, and an active circulation of atmospheric water was undoubtedly established long before the quicksilver deposits were formed.

7. Cretaceous or Later Epochs of Copper Concentration in Sedimentary Rocks.—In addition to these five classes, whose connections with igneous rocks are indubitable, the disseminated copper ores of the southwest should find mention. They occur in sandstones, shales, or conglomerates ranging from the Carboniferous to the Cretaceous, and, in most cases, chalcocite is the original ore; frequently small amounts of silver are present. New Mexico, Arizona, Utah, Colorado and Texas offer numerous examples of this class. While their origin is not wholly clear, many observers believe that they represent concentration by ground waters of small amounts of copper originally derived from the pre-Cambrian deposits and since distributed through late sedimentary beds. In similar deposits we find ores of uranium, vanadium and radium.

Summing up we have then in the Cordilleran region:

	Principal metals	Principal rocks associated
1. Deposits of the Pre-Cambrian period:	Gold and copper.	Granites. Diorites, gabbro.
2. Deposits of the early Mesozoic epoch:	Copper.	Basalt, diabase. Gabbro.
3. Deposits of the late Mesozoic epoch:	Gold.	Granodiorite. Quartz-monzonite, gabbro.
4. Deposits of the early Tertiary epoch:	Gold, silver, copper, lead, zinc.	Granodiorite. Quartz-monzonite. Monzonite, with corresponding porphyritic rocks.
5. Deposits of the late Tertiary epoch:	Gold, silver.	Andesite, latite. Rhyolite.
6. Deposits of the Post-Pliocene epoch:	Quicksilver.	Basalt, andesite.
7. Cretaceous or later concentrations in sedimentary rocks.	Copper, vanadium, uranium, radium.	Sandstone, shale, conglomerate.

In deposits that are clearly connected with igneous rocks metallization is certainly a function of varying pressure and temperature; these factors being dependent upon depths below the surface and other conditions; metallization is also dependent on the nature of the rocks in which deposition takes place. Primarily, however, it is probably a consequence of magmatic differentiation.

It is well established that magmas of different types contain different associations of the rare metals. For instance, tin and tungsten characterize acidic rocks, while nickel and cobalt are found chiefly in magmas rich in ferro-magnesian constituents. At the same time rocks of a given general composition may, in different localities, vary considerably in the quantity of rarer metals contained.

Our knowledge of the content of rare metals in igneous rocks is fragmentary, but it is at least often the case that ore deposits formed after the eruption of an igneous rock will contain the rarer metals characteristic of it. It may be true that for each differentiated rock type there are corresponding types of deposits, varying with the conditions of deposition.

As periods of long continued differentiation may materially modify the composition of a magma, it is conceivable that this might find expression in a progressive change in the character of ore deposits successively formed during these periods. The quicksilver deposits of California, for instance, may be the ultimate result of such long continued differentiation.

As we look back over the wide domain of mineral deposits we perceive the strong tendency to concentration of common or rare elements, by magmatic differentiation, by solution or by mechanical transportation; we perceive also cycles of transformations, based on the laws of stability of chemical compounds. Even when deeply buried the deposits may suffer many changes. Near the surface they are constantly subject to transmutations involving both concentration and dispersion. A few stable compounds are formed while the rest are of the elements scattered by mechanical and chemical transportation. Some constituents are carried down into the earth by the underground circulation of water, perhaps to form new deposits in other rocks. Ultimately erosion sweeps away the wreckage of the old deposits into basins of sedimentation where the elements may be reconcentrated, be it by direct precipitation or by the aid of living matter. The sediments may again be lifted and corrugated, again destroyed by erosion and new eras of concentration begin.

In one aspect the science of mineral deposits is frankly utilitarian, but from the viewpoint of pure knowledge it records the principles governing the cycles of concentration of the elements. It traces the processes by which the primeval gases and magmas have become differentiated into the manifold complexity of the earth's crust.

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