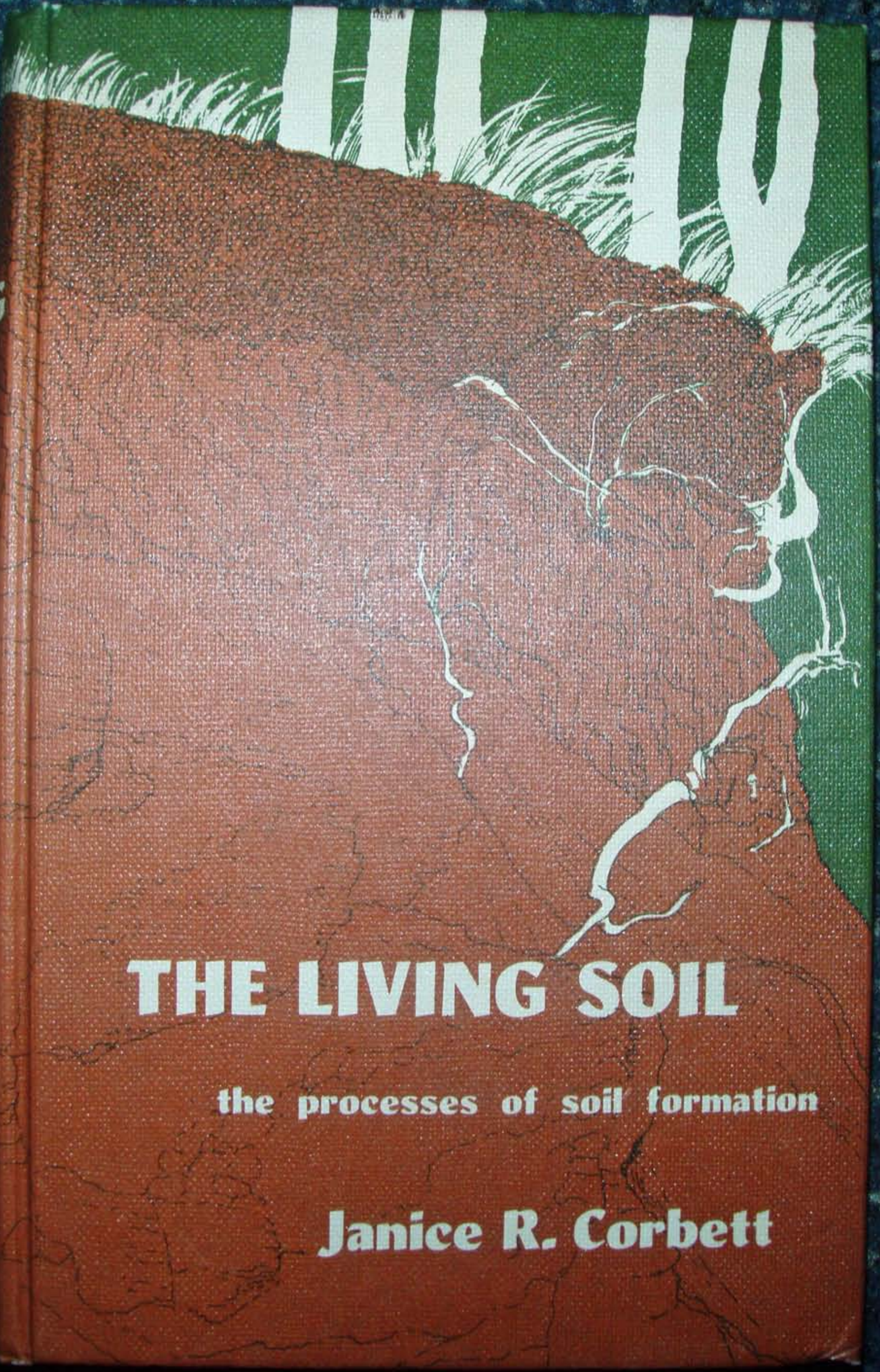

THE LIVING SOIL

the processes of soil formation

Janice R. Corbett

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The Processes of Soil Formation

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Foreword

The Living Soil has been written by Dr Janice Corbett at the request of the Council of the Geography Teachers' Association of New South Wales in order to provide teachers, university students and senior students in high schools with basic information which will enable them to study soils in the field and will assist them to understand the more specialised textbooks and journal articles written about soils.

Dr Corbett was selected as author because she recently completed her Doctorate of Philosophy thesis in the School of Geography, University of Sydney, on the pedology and mineralogy of soils near Inverell, N.S.W., and because she has generously shared her knowledge with many teachers, both at refresher courses organised at Teachers' Colleges and Universities and during field days with senior students. From such activities she has been able to assess the deficiencies in her audiences' knowledge of pedology, while her research work has revealed to her the difficulties encountered by teachers and students who need ready access to up-to-date information on the soil types of Australia. In pedology, as in geomorphology and climatology, research workers have become more and more critical of the information supplied in physical geography textbooks. Their criticisms are substantial: that description is emphasised while the soil-forming processes are relatively neglected; that soil profile models are often based on northern hemisphere examples rather than on Australian field examples; and that the various soil classifications are not examined critically nor their suitability assessed for Australian conditions.

Dr Corbett has organised *The Living Soil* so that there is a sequential development of concepts and skills associated with the study of soils in the field. Her early chapters progress from the more elementary soil-forming processes interpreted from field soil properties to the problems of soil classification; then there are five chapters which discuss selected Great Soil Groups classified according to the principal formal processes of soil formation; finally, there is a study of relict soils in Australia. A comprehensive Glossary of terms has been added, and this should be most helpful to students.

In Australia today, awareness of the value of a scientific approach to the soil is rising rapidly. Not only is soil study a major research frontier, but schools, colleges and universities are devoting more and more time to equipping students with a theoretical and practical grasp of the subject. Symptomatic of the trend is the fact that—as this book goes to press—schools in New South Wales are, for the first time, being issued from government sources with a testing kit for the study of soil in the field. (Dr Corbett's second chapter describes how to use such a kit.)

Dr Corbett's book therefore makes a timely as well as an informative contribution; indeed, it may be regarded as the most up-to-date foundation study of soils now available. Very usefully, it has been able to draw on the proceedings of the 9th International Soil Congress, held in Adelaide in 1968. While the book is sure to develop interest in soil-study among students of geography, geology and agriculture, it should also assist intelligent laymen in various fields, such as farming and conservation.

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My interest in soils was first stimulated by the lectures of Associate Professor T. Langford-Smith of the School of Geography, University of Sydney, who also greatly assisted me in my doctoral studies. Professor J. A. Mabbutt of the School of Geography, University of New South Wales, has been very helpful in his encouragement to finish the book despite other pressures. And my husband, John, has helped in a hundred ways apart from the technical ones of instructing me in the laboratory analysis of soils and augering most of the soil profiles I have studied in the field.

Finally, I would like to thank the many Geography teachers who have discussed the planning of the book at conferences and field days. Their enthusiasm has been an inspiration. In particular I shall welcome any suggestions from them and shall try to incorporate these if the book reaches a second edition.

J. R. C.

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FROM THE EARLIEST TIMES, MAN'S SURVIVAL HAS DEPENDED upon the soil. Even before he learned the art of cultivation, man was closely related to the soil through hunting and gathering. These pursuits could only succeed where vegetation—and therefore soil—was present. Without soil the world would at best be covered with mosses and lichens, which could not support either man or the animals that man preys on.

Soil and vegetation have evolved together. When a fresh rock surface such as a lava flow weathers, soil and vegetation develop in a finely adjusted relationship. It is a matter of common observation that vegetation changes as the depth and character of the soil change, even within restricted localities. Areas of shallow, stony soils can produce only poor quality, low yielding crops. In Australia there are many areas where inferior soils produce low crop yields and poor pastures, but fortunately there are also areas where fertile soils enable better yields to be obtained. Animal life too is deeply influenced by the quality of the soil: cattle fed on pastures where the soil is deficient in copper soon become unhealthy, and sheep grow a 'steely' wool that lacks crimp.

THE DEVELOPMENT OF SOIL SCIENCE

Early scientists paid little attention to soil. They were concerned with what seemed to them more interesting matters, such as mathematics and astronomy, and they left the soil to the peasant. But the illiterate peasant could not analyse and study the soil; all he could do was pass on his small knowledge of the art of cultivation from generation to generation. Soil study was also hindered because there was little travel in earlier times.

This prevented regional comparison of soils; moreover, such comparison was all the more difficult because prolonged, repeated cultivation of farms destroyed many of the original soil features. The Roman scientist, Columella, dealt with soils at length in his famous book on Husbandry, compiled in 50 A.D., but he was concerned with practical problems of soil management and productivity, not with an independent general study of soils; and his work was restricted to the soils of the Italian Peninsula.

In Western Europe, at the time of the French Revolution, attention was again directed towards soil fertility. This was partly prompted by the fact that population was growing more rapidly than food supply. Chemical analyses of crops revealed that they extracted certain elements from the soil. The growing of plants not in soil but in water containing nutrients (the study of this is now called hydroponics) revealed that many elements were needed by plants during their growth. Because of these studies, the soil came to be regarded as a storehouse for plant nutrients, the plants extracting the nutrients as required, thus depleting the soil. This idea, though widely accepted at the time, failed to explain the differences inherent in soils in the first place—why some soils were productive and others were not. It also failed to explain why certain soils, especially the alluvials, could be cultivated continuously and yet remained fertile, while others became unproductive after just a few crops. The soil was soon found to be too complex for a simple 'balance sheet' theory; moreover, early soil studies in Europe tended to pay too much attention to plant responses at the expense of study of the soil itself.

It is perhaps not surprising that the study of soil as a natural body began in Russia where the emphasis was on the development of vast expanses of new territory exhibiting diverse vegetation and climate. It was possible for Russian scientists to examine soils in their original state because of the existence of great areas of virgin soil (soil that has not been ploughed). They noted that soils are made up of *horizons* which are related to one another and to the underlying parent material. Previously each soil horizon was regarded as a separate geological stratum. From these studies of the soil as a natural body evolved a new science called Pedology, the study of the morphology and genesis of soil.

V. V. Dokuchaev (1846-1903), who is often referred to as the 'Father of Pedology', concentrated on the relationship between the soil horizons and their environment (this is referred to as the *genetic theory* of soil formation). In his famous book, *The Russian Chernozem*, published in 1883, he noted that the chernozem soil (a heavy black soil found in the steppe areas)

was always associated with grassland vegetation. He saw that the chernozem is not a separate geological stratum, as was previously believed, but is found under grassland on many parent materials, between narrow limits of precipitation and evaporation. He considered that the chernozem and its grassland were intimately related. Dokuchaev also studied the soils of the great coniferous forests of Russia and was able to explain the peculiar ash-grey, bleached layer immediately below the dark mineral layer in the forest soils. He related this layer to those above and below it, to the moist climate and to the forest vegetation. He was thus able to account for it not as a separate entity but as an integral part of a sequence of layers which constitute a soil termed a *podzol* (from a Russian word meaning 'ash-grey,' 'ashes underneath'). The podzol and the chernozem clearly illustrated the role of climate in soil formation; this led later Russian pedologists to emphasise the *zonal* theory of soils which relates soil zones to climatic zones. Whilst they stressed the role of climate in accounting for soil diversity, they also recognised that there were soils which, because of some peculiar geological stratum or vegetation or drainage condition, did not develop normally for their climatic zone. These they called *intrazonal* soils.

The Russian pedological works were not translated into English for about thirty years. During this period, soil studies in the United States were developing independently. They had a strong geological bias since most of the work was done by the United States Geological Survey: at the turn of the century, American geological maps sufficed for soil maps, and the role of climate in the development of soils received little attention. C. F. Marbut (1863-1935) translated some of the Russian pedological studies in 1914, but he considered both the climatic (zonal) and geological classifications of soils inadequate as he believed soil to be a complex natural body evolved from underlying parent rock but different from it because of the influence of climate and organisms. He contended that only soil properties could provide a satisfactory basis for the classification and study of soils. His classification, the first based solely on soil properties instead of on climate or geology, was presented at the First International Conference in Soil Science, in 1927.

Pedology, then, relative to sciences such as mathematics and chemistry, is a young science. Its province, the genesis and morphology of soil, forms only a small part of the broader field called Soil Science which embraces all aspects of the study of soil. Agricultural students study the general field of soil science, including pedology, soil physics (the study of the soil's physical properties such as pore space, water movements, density, temperature), soil chemistry (the study of the soil's chemical properties such as the movement of ions through the soil, mineral weathering, the composition of the soil's water, air and mineral matter), soil microbiology (the study of the soil's living population, both flora and fauna), soil fertility and plant nutrition (the study of the plant's response to soil) and soil mineralogy (the study of soils under the microscope). These fields rely heavily on each other; for instance, a pedologist may use soil chemical or mineralogical techniques to interpret soil morphology, or a soil chemist may need to know a soil's morphology. This book is mainly concerned with pedology, which is sometimes referred to as 'the study of soils in the field'. Since in pedology soils are related to their environment of formation, this aspect of soil science is the most useful for geographers. A knowledge of pedology is basic to many branches of geography, especially physical geography.

DEFINITION OF SOIL

Advance in pedology depended on the realisation that soil is a natural body formed in response to atmospheric forces but different from its parent material. So, to the pedologist, soil is quite different from many other weathered deposits covering the earth's surface. Agriculturists, on the other hand, tend to regard soil merely as the medium for plant growth. But plants will grow in water containing nutrients, in sand dunes, in stream deposits (alluvium), and in dust deposited by the wind (loess). Are all these to be regarded as soil? They certainly support plant life but they have none of the features which directed the attention of the early pedologists to soil as a natural body. The essential features of soil, considered as a natural body, are the *horizons* which reflect the action of atmospheric forces on weathered material. Mosses will grow on bare rock surfaces, but if exposure to the atmosphere is long enough, the rock will weather, vegetation will become more complex, and eventually soil horizons will develop. Similarly, if loess, dune sands or alluvium remain long enough as stable surface deposits, they too will become differentiated into soil horizons. These transformations into horizons in response to atmospheric forces are referred to as the *soil-forming processes*. They make soils unique from all other weathered materials.

Weathered deposits such as alluvium may also contain layers, but these are not the result of atmospheric forces and arise from sedimentation or deposition. Layers in alluvium, unlike the horizons of soils, may be quite unrelated to each other and independent of their place of deposition. One layer may contain sand and pebbles, while the layer on top may contain silt and clay. These layers are related to their source material rather than to

their place of deposition. Soil horizons are only partly related to their parent material: many properties are more closely affiliated with the environment of formation. Pedologists refer to the layers in soils as *horizons* to distinguish them from the layers of other deposits. Soil horizons are related to each other and to the underlying parent material, which may be rock, alluvium, loess or any other non-consolidated deposit.

The soil-forming processes which cause this transformation into horizons are many and varied. Certain processes are common to many soils (for instance, the decomposition of organic matter and its incorporation into the profile) while others are restricted to certain soil types (for instance, the leaching of silica in lateritic soils). The removal of clay particles from the topsoil to the subsoil is a soil-forming process which is found in many soils under diverse climates. This removal of clay leaves the topsoil with a sandy texture, while the subsoil becomes enriched with clay. Thus the sandy nature of the topsoil is directly related to the clayey nature of the subsoil, and the amount of clay present in either horizon is in part controlled by composition of the parent material. The horizons constitute a unit, so that material removed from one horizon may be deposited into another horizon either above or below it. The mobilisation of clay presupposes a weathered parent material and so could occur as soon as alluvium or loess become stabilised. The mobilisation of constituents from the topsoil and their deposition into the subsoil causes alluvium to be differentiated into horizons, and once this has occurred, the upper part of the alluvium may be referred to as *soil*. The layers too deep to be affected by the soil-forming processes are still termed *alluvium* and may continue for many feet in depth.

In the case of fresh rock, the mobilisation of clay cannot proceed until the primary rock minerals have changed chemically into clay-sized particles, so weathering operates concurrently with the processes causing horizon differentiation. In fact, many pedologists include physical and chemical weathering in their list of soil-forming processes.

In order to distinguish soil from all other weathered materials such as those described above, a definition must be chosen that will include the horizons which develop in the course of the soil-forming processes. One such definition is that of Joffe (1949) who regards soil *as weathered material in genetically related horizons*; this indicates that the soil horizons form together, distinguishing them from other layers, which may be deposited quite independently, in other weathered materials such as alluvium.

The role of plants in the genesis of soil horizons is vital. Some pedologists even claim that horizon formation is dependent on the growth of vegetation, and that without it, a soil will not form. This is certainly true in dune sands which will not develop a soil profile until vegetation stabilises them, enabling the upper layers to become darkened by organic matter. However, in theory, one could envisage the translocation of clay from the topsoil to the subsoil without the presence of organic matter. Such cases would, of course, be unlikely to occur in nature, as once a surface remains stable and water is available vegetation soon moves in. The definition of soil may therefore be modified to include the role of plants: *soil is weathered mineral and organic matter found in genetically related horizons in response to subaerial processes.*

SOIL HORIZONS AND THE SOIL PROFILE

The sequence of horizons down to and including the parent material is referred to as a *soil profile*. This term is generally restricted to horizons which are known to be all part of a single soil. Where a sequence of older, buried soils is overlain by the modern active soil, the sequence may be referred to as a series of soil profiles.

The soil profile consists of two parts, the solum and the weathered parent material. The *solum* is the horizons which have been changed from the parent material by the processes of soil formation: they therefore have their own organisation and differ markedly from the parent material. Beneath the solum there is usually a zone of *weathered parent material* which is too deep to be transformed into solum horizons, but which is weathered. This zone of weathered parent material is found even in soils derived from the underlying rock and is often much deeper than the solum. Since the weathered parent material is not sorted into horizons, it is more like the fresh rock than the solum in its properties. The whole mantle of weathering from the top of the solum to the fresh rock is referred to as *regolith*. These features are shown in Fig. 1.1.

The naming of the horizons of the soil profile is one of the most important tasks of the pedologist: it requires a thorough understanding of the processes of soil formation and the relationships between the soil horizons. Many systems for the designation of soil horizons are in use. In most of these, the letters used for the horizons relate to the movements of constituents through the profile and therefore stress the genetic relationship between the horizons.

In most systems the first distinction to be made is between horizons that have lost constituents and horizons that have gained constituents: the former are termed *eluvial* horizons from Latin words meaning 'washed out of' and the latter are termed *illuvial* horizons from Latin words meaning 'washed into'. The removal of constituents from the upper soil horizons is referred to as eluviation and the deposition of some of the eluviated constituents into lower soil horizons is referred to as illuviation. Most systems of designating soil horizons use the letter *A* for the eluvial horizons and the letter *B* for the illuvial horizons. Clay, iron, humus and lime are the most commonly eluviated and illuviated constituents so that the *A* and *B* horizons may be subdivided on the basis of properties which reflect the movements of these constituents. Many systems of suffixes are used for the major horizons—some countries use numbers, others letters, and others both numbers and letters, resulting in a great deal of confusion in soils literature.

The Food and Agriculture Organization of the United Nations and the United Nations Educational, Scientific & Cultural Organization (referred to hereafter as FAO and UNESCO) which are preparing maps of world soils, and the International Society of Soil Science have for some time been concerned about the confusion caused by the various systems used for designating soil horizons and the resultant difficulty in communication between pedologists.

A proposal for an international system of designating soil horizons was put forward in 1966 and modified in 1967. The modification was discussed at the 9th International Conference in Soil Science (Adelaide, August 1968). Yet to be finalised, this system is indeed a simple one and, like most other systems, it depends greatly on the movement of constituents through the soil profile and uses the letters *A* and *B* for the eluvial and illuvial horizons respectively. However, additional letters are used for the major soil horizons: these letters are abbreviations of meaningful words and so are easy to remember.

The letters proposed for the major horizons in the 1968 draft of the International system are as follows, from the top of the profile down (but G can occur anywhere and K can occur in *B* or *C* horizons):

O	horizons consisting mainly of organic matter[1] from the vegetation; accumulates under conditions of free aeration
A	eluvial horizons consisting mainly of mineral matter mixed with some humified organic matter[2]

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[1] The organic matter content must be more than 30% if the soil is clayey and 20% if the soil is sandy.

[2] The organic matter content must be less than 30% if the soil is clayey and 20% if the soil is sandy.

E	strongly eluviated horizons having much less organic matter and/or iron and/or clay than the horizons underneath; usually pale coloured and high in quartz
B	illuvial horizons characterised by concentrations in clay, iron or organic matter, alone or in combination; some lime may accumulate, but if the accumulation is excessive, the letter *K* is used
K	horizons containing appreciable[3] carbonate (usually mainly lime or calcium carbonate)
G	gleyed horizons which form under reducing conditions with impeded aeration, reflected in bluish, greenish or greyish colours
C	weathered parent material lacking the properties of the solum and more like the fresh parent material
R	consolidated bedrock.

Horizons which are transitional between the major horizons may be designated by both the letters of the major horizons, the first letter marking the major horizon to which the transitional horizon is most like, e.g., *EB* or *BE*. Mixed horizons may be designated by both the letters of the major horizons, separated by a stroke, e.g., *E/B, B/C*.

Numbers or letters, or both, can be used as suffixes to the major horizons. The pedologist may merely divide the horizons using numbers in continuous sequence from the top down, e.g., *A1, A2, B1, B2, B3*. If, however, the pedologist understands the genetic processes operating within the soil, he may use small letter suffixes which are abbreviations of meaningful words such as humus (*h*) and fermented (*f*), e.g., *Of, Oh*. If the pedologist is unsure of the genesis of every horizon or if there is more than one horizon with the same major letter and suffix, both numbers and letters can be used, e.g., *A1h, A2h, B1fe, B2fe*. In this book, where possible, letter suffixes are used, rather than numerical suffixes, in order to stress the genetic significance of each horizon.

The proposed suffixes for the International system of horizon designation are:

for the accumulation of organic matter:

l	litter: fresh accumulation of leaf litter	applied to *O* horizons	accumulates under free drainage and aeration
f	fermented, partly decomposed organic litter		
h	humus, well decomposed organic matter (can be applied to *O, A* or *B* horizons)		

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[3] The carbonate content must be at least 50% before the letter *K* is used, otherwise the suffix *ca* (e.g., *Bca*) is used.

a	(from German *anmoor*, peaty) well decomposed organic matter (applied to *A* horizons)	accumulates under impeded drainage and aeration, i.e., hydromorphic conditions
o	poorly decomposed organic matter (applied to any horizon)	

for the accumulation of sesquioxide (the hydrated oxides of iron and aluminium):

cn	accumulation of concretions or nodules of sesquioxides
fe	illuvial accumulation of iron
ox	residual accumulation of sesquioxides
r	concretionary or gravelly layers

for salt accumulation:

ca	accumulation of calcium carbonate
cs	accumulation of calcium sulphate
na	(from Latin *natrium*, sodium) high percentage of sodium on the clay (usually detected by a high *pH*—see Chapter 2)

for clay accumulation:

t	(from German *ton*, clay) illuvial accumulation of clay
v	(from German *verwitterung*, weathering) accumulation of clay by alteration *in situ*

other suffixes:

b	buried; applied to any buried horizons
g	gleying; strong mottling reflecting variations in oxidation and reduction as a result of periodic wetness: not enough reduction to warrant use of major letter *G*
m	strong cementation or induration
p	disturbed by ploughing or other tillage practices
x	fragipan (a hardpan cemented by colloids).

A typical sequence of soil horizons which could be found in podzolised soils is shown in Fig. 1.1, which illustrates the International system of horizon designation.

REGOLITH	SOLUM	organic mat	Ol	fresh leaf litter
			Of	partly decomposed leaf litter
			Oh	well decomposed leaf litter
			OA	transition between organic mat and first mineral horizon
		eluvial horizons	Ah	weakly leached horizon with humus accumulation
			E	strongly leached horizon, light in colour
		illuvial horizons	Bh	strong humus enrichment
			Bfe	strong iron enrichment
			Bfe, g	strong iron enrichment with mottling due to periodic wetness
			CB	transition between illuvial horizons of solum and weathering rock
		weathered parent material	C	altered rock, with rock structure preserved
		bedrock	R	fresh rock

Figure 1.1: The soil horizons. (The designations are those of the Proposed International System.)

The degree of subdivision of a soil into horizons, together with the use of suffixes, depends to a certain extent on the eye and the tools of the pedologist. In the field, a careful description of the soil profile may reveal three or four horizons, distinguished mainly by, say, colour and structure. Laboratory analyses of samples from the profile may reveal differences in chemical composition within some of these horizons and so lead to further subdivision. If the proposed International system is used, additional suffixes may be used without any change in the major letters.

THE SOIL-FORMING PROCESSES

Essential to the study of the soil is an understanding of the processes which bring about horizon differentiation. It should be possible to recognise some of the soil-forming processes when you meet them in the field. Try to realise at the outset that the soil is a *dynamic* body and that these processes are going on all the time. If they ceased, the soil horizons would soon lose their character and the profile would degenerate, because new materials are constantly being added to the soil. For example, if the process of humification ceased, the organic matter added by plant decay would soon 'choke' the profile, preventing the movement of the other mobile constituents and forming a peat rather than a soil. Careful profile description should reveal the action of these processes, for it is the depletion and accumulation of constituents that the pedologist seeks when he describes a soil profile.

The soil-forming processes may be considered at various levels, from the apparently simple, single processes to the more complex, multiple processes. Although these processes provide the theme of the chapters on the Great Soil Groups (Chapters 7 to 12), some of the single processes are mentioned here to demonstrate the dynamic nature of soil. The soil-forming processes are the reason for the title of this book, *The Living Soil*. The more formal processes, such as laterisation or podzolisation, are described in later chapters in their appropriate Great Soil Groups (lateritic soils and podzols, respectively). These processes, which are select combinations of several single processes are restricted to a few soil types, whereas each single process covers a wide range of soil types. Thus the formal processes may be regarded as a combination of many single processes which can be enumerated separately, but which occur concurrently and sometimes even in combination in the soil. However, in the field, these may be seen as single, separate processes in the soil. An example to illustrate the various levels of complexity of soil-forming processes can be taken from the illuviation of constituents in podzolic soils.

Dark brown and black colours in soils usually originate from organic compounds, and red and yellow colours from iron compounds. The *Bh* horizon of many podzolic soils is dark brown or black indicating illuviation of humus from the *A* and *E* horizons, while the *Bfe* horizon is often reddish-yellow indicating illuviation of iron from the *A* and *E* horizons. The *Bh* and *Bfe* horizons may also contain appreciable quantities of clay illuviated from the *A* and *E* horizons. One might be tempted to consider the illuviation of iron and humus as separate processes, but it has long been known in pedology that iron and humus are frequently mobilised and deposited together. Laboratory analyses of *Bh* horizons of podzolic soils usually reveal quite high concentrations of iron as well as humus, but in the field the dark colour of the humus masks the lighter colour of the iron. So, though profile description may reveal apparently single processes, the student must realise that they probably operate in conjunction with other processes; for example, in podzolic soils the bases are depleted at the same time as the translocation of the iron, clay and humus.

What are some of the soil-forming processes common to many soil types? The most universal and fundamental process is, of course, weathering. This is the alteration of rock minerals into smaller particles, sometimes with different structure and chemical composition. Weathering happens wherever rocks and rock particles are exposed to the atmosphere; it is common to soil and non-soil materials alike. The term 'weathering' is applied to many and varied processes. These are discussed in Chapter 3 which deals with the soil's mineral matter. At this stage, it should be noted that weathering provides the open or porous structures through which water can penetrate to remove the mobile constituents of the soil, some of which are directly the result of weathering in their present position during soil formation. Weathering and horizon differentiation proceed concurrently and they are closely inter-related; sometimes the rate of soil formation depends on the rate of weathering, especially in soils derived from fresh rock. Many textbooks, however, separate weathering and horizon differentiation. They do so in order to simplify the description of these processes; but remember that in soils derived from rocks differentiation into horizons cannot proceed without weathering. In soils from unconsolidated parent materials, weathering accompanies soil formation, even though the parent material is pre-weathered.

Perhaps the most important processes causing horizon differentiation are those grouped under the term *leaching*. Leaching operates to some extent in all soils and is fundamental to all the more complex, multiple soil-forming processes such as laterisation, podzolisation and calcification. It means the removal of constituents from the soil by water, and is therefore most effective in humid areas and least effective in dry ones. Various types of leaching appear as single processes, and the leaching of individual constituents can be detected in the field.

Soil constituents offer varied resistances to leaching. Some elements, such as silicon, resist leaching quite well under certain conditions, while others, such as sodium, are readily removed under all but the driest conditions. It is fortunate that the elements toxic to plants are those that leach most readily from the soil. If some of the salts (especially sodium chloride) did not easily leach from common rock constituents, plant growth would not be possible in many soils because even small concentrations of some salts are toxic.

But not all leaching is beneficial. Excessive leaching robs many soils of essential plant nutrients which can only be extracted by plants in the soluble form—the very form in which they are leached from the soil. Huge quantities of nitrogenous fertilisers are used every year by farmers in order to offset the natural wastage of soil nitrates by leaching. On the other hand, there are soils that would be all the better for greater leaching of certain elements. For example, in tropical red soils, excessive iron and aluminium form complexes with soil phosphates, causing a deficiency in the amount of soluble phosphorus available for plants. The productivity of these soils would increase if there were less iron and aluminium, though these elements, unlike salt, are

not toxic. Leaching may arbitrarily be divided into two types depending on whether or not the leached constituents are redeposited in the soil profile. The first type involves the complete removal of the constituents from the soil by drainage water, while the second involves the removal of constituents from some horizons and their deposition in others, that is, eluviation and illuviation.

Eluviation and illuviation are not the only movements of soil constituents through the profile: in some soils, especially where there is a marked dry season, constituents move up the profile through the soil capillaries. For example, salt may concentrate at or near the surface of desert soils by capillary rise in a long dry season. The indurated zone of lateritic soils (described in Chapter 7) results partly from the capillary rise of iron and aluminium from the lower horizons in the profile and their deposition near the surface as hydrated oxides when the soil dries.

The constituents which are leached from soils may be divided into ions and colloids according to their size and solubility. *Ions*, which are positively (*cations*) or negatively (*anions*) charged bodies, exist only in the soil solution, and so their concentration depends to a certain extent on the solubility of the source material. Ions, which may form from elements (e.g., the sodium ion, Na^+) or the molecules (e.g., the ammonium ion, NH_4^+) are extremely small. *Colloids* are larger than ions, but are also quite small, so small in fact that they are governed by their enormous surface area[4] and may be carried long distances in suspension (as distinct from solution in the case of ions). Colloids, like ions, may move through the soil from horizon to horizon, but they are less likely to be completely removed from the soil profile than are ions, many of which find their way into drainage waters.

Ions may be released during the weathering of primary rock minerals; for example, sodium (Na^+) and calcium (Ca^{2+}) derive from feldspars. They may form during the humification process; for example, ammonium (NH_4^+) and phosphate (e.g., $H_2PO_4^-$). They may also originate from gases dissolved in the soil; for example, the carbonate ion (CO_3^{2-}) derives from dissolved carbon dioxide. Ions may leave the soil in drainage waters or adhere to the soil colloids (see Chapter 3). The ease of removal of ions from the soil depends firstly on the solubility of compounds from which they are derived, and secondly on the nature of other ions with which they could combine. Sodium is easily removed because most sodium compounds such as sodium chloride and

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[4] A cube of material cut into particles 1 mm across (sand size) has a surface area of 60 sq. cm; but cut into particles 1/10,000 mm across (clay size) has a surface area of 600,000 sq. cm.

sodium nitrate are soluble, but although some calcium compounds are soluble (e.g., calcium bicarbonate and nitrate) others have very low solubilities (e.g., calcium carbonate and sulphate).

An elementary knowledge of chemistry, especially of the solubilities of the various common constituents of soil, will enable you to predict the ease of removal of soil constituents. If you do not have this knowledge, the following generalisations may help; but bear in mind that the removal of elements from the soil depends a great deal on the conditions within the soil. The alkali cations (sodium and potassium) are the most readily leached of the common soil constituents. Next are the alkaline earths (calcium and magnesium). These two groups are very thoroughly leached from the soils of wet regions (lateritic and podzolic soils). Silicon compounds are fairly insoluble, but under certain conditions such as those existing in lateritic soils, silicon may be leached as silicic acid. The elements most resistant to leaching are iron, aluminium, manganese and titanium; although they are readily supplied in ionic form by the weathering of many primary minerals (e.g., olivine, magnetite), they form very insoluble compounds with oxygen and water. These compounds are termed the *sesquioxides* and are found in most soils, but they are highly concentrated in lateritic soils (soils of hot, wet regions) where most of the other constituents have been leached. The order of removal may be generalised as:

$$\left.\begin{matrix}\mathrm{Na}\\ \mathrm{K}\end{matrix}\right\}\text{greater than}\quad\left.\begin{matrix}\mathrm{Ca}\\ \mathrm{Mg}\end{matrix}\right\}\text{greater than}\quad\mathrm{Si}\ \text{greater than}\quad\left.\begin{matrix}\mathrm{Fe}\ \text{and}\ \mathrm{Mn}\\ \mathrm{Al}\qquad\mathrm{Ti}\end{matrix}\right\}$$

Thus the soils of very wet, hot regions are dominated by the sesquioxides. The soils of humid, cooler regions, where there is less intensive leaching, contain a great deal of silicon compounds, and some sesquioxides, but few bases (the alkali and alkaline earth cations). The soils of semi-arid regions contain considerable calcium and magnesium compounds but not a great deal of sodium and potassium, while the soils of arid regions often have a 'salting' problem—that is, too much sodium.

Not all of the ions in the soil adhere to the colloids. Many combine to form deposits lower in the soil profile; for example, lime (calcium carbonate) is often found in the *B* horizons of semi-arid soils. These deposits, which are often whitish, are easily detected in the field, whereas actual ions are not easily traced.

The mobile colloids which are translocated through the soil are also easy to detect. Silicate clays and organic colloids (see Chapter 3) are carried in suspension through the soil to be deposited in other horizons. Silicate clays appear as shiny glistening surfaces on the soil aggregates in the horizons in which they are deposited. These shiny surfaces, very prominent in the field, are termed clay skins. Not all clay skins arise from the deposition of constituents eluviated from upper horizons. Some form from the deposition of constituents within the horizon from the soil solution as the water dries. Clay skins are usually quite conspicuous in wet soils, especially when they differ in colour from the host soil material. Silicate clay skins are usually coloured by iron deposited with the silicate clay. Organic skins are black and dull, whereas manganese skins are usually black but more lustrous.

Other soil-forming processes will be mentioned later. It is the object of this book to demonstrate how these processes may be detected in the field and how they may be used to divide the world's soil spectrum into manageable categories. Chapter 2 will tell how to detect the processes listed above as well as other processes, but some examples are given here to show the significance of the various soil properties described in the field.

Soil acidity or alkalinity (soil reaction) indicates a great deal about the removal of cations from the soil. An acid soil is depleted of the basic cations which must have been leached, while a moderately alkaline soil probably contains calcium carbonate. The application of hydrochloric acid to basic soils will tell whether or not lime is present. A strongly alkaline soil contains sodium and potassium compounds, though some soils which are close to neutral in reaction may also contain these.

The colour changes observable in the soil profile indicate a great deal about the mobilisation of constituents, especially organic matter (black), iron (red and yellow) and lime (white). Variations in soil texture are largely a consequence of the mobility of clay within the soil.

Of course, it takes experience to immediately recognise all the properties in a given soil profile. The only way to gain this experience is by systematically describing as many soil profiles as possible. The following chapter tells how to approach a soil profile, what to look for, and the significance of the various properties. When approaching a soil, realise that it is a dynamic, not a static body, and that many processes have operated to make it the way it is. Rather than adopt the morphologic approach of describing for description's sake, the best course is to adopt the dynamic approach of describing the soil in order to find out how it formed.

REFERENCES AND FURTHER READING

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Chapter Two

Description of Soil in the Field

CONTENTS

FIELD EXCURSIONS ARE AMONG THE MOST INTERESTING ACTIVITIES of any course. The study of soils is an easy and pleasant part of field work, especially in geography. *Any* soil can prove rewarding, the field techniques are the same for all soil profiles, and very little equipment is needed to produce first-rate results.

In a local area, it is true, there is only a small part of the world's soil spectrum, but through the study of local soil types the student can gain an appreciation of the world's Great Soil Groups, which are described later in this book and also in most physical geography textbooks. This appreciation can be gained by observing the local variations which occur in soils and relating these to the system of soil classification given in this book.

The soil classifications most useful to pedologists are based on soil properties as seen in the field. This is why it is important to become familiar with systematic profile description. Also, the study of a number of soil profiles will enable the student to steadily gain a better understanding of the soil-forming processes which are basic to pedology.

THE STUDY OF SOIL IN THE FIELD

The study of soils may involve systematic profile description at a soil site readily accessible to students; or each student may be set an assignment to carry out on his own, in which case this book should be closely followed to make sure that no important soil property is overlooked. Students are well advised to begin early in their courses to look for well-exposed soil profiles because they are seldom easy to find. Building excavations, erosion gullies, road and railway cuttings and quarries usually provide the most

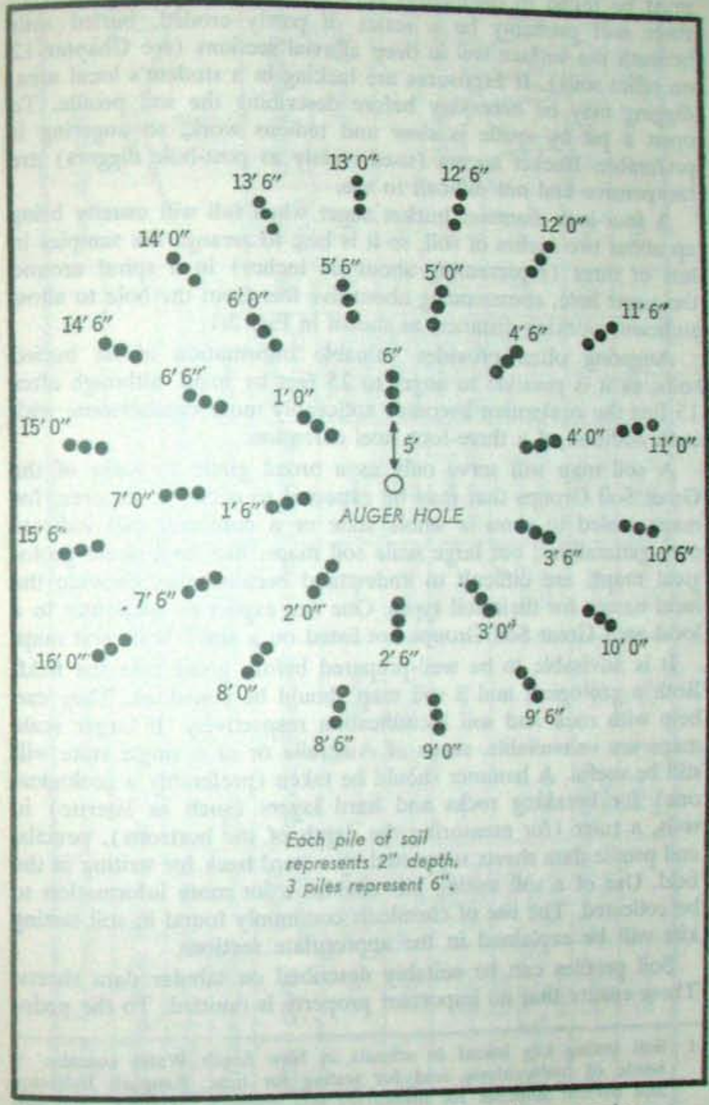

Figure 2.1: Method of placing samples when augering.

suitable exposures. Some creek sections are suitable, but care must be taken to distinguish between soil and alluvium; indeed, there will probably be a series of partly eroded, buried soils beneath the surface soil in deep alluvial sections (see Chapter 12 on relict soils). If exposures are lacking in a student's local area, digging may be necessary before describing the soil profile. To open a pit by spade is slow and tedious work, so augering is preferable. Bucket augers (used widely as post-hole diggers) are inexpensive and not difficult to use.

A four-inch diameter bucket auger when full will usually bring up about two inches of soil, so it is best to arrange the samples in lots of three (representing about six inches) in a spiral around the auger hole, commencing about five feet from the hole to allow sufficient working distance, as shown in Fig. 2.1.

Augering often provides valuable information about buried soils, as it is possible to auger to 25 feet by hand, although after 15 feet the equipment becomes noticeably more cumbersome with each addition of a three-foot steel extension.

A soil map will serve only as a broad guide to some of the Great Soil Groups that may be expected to occur in an area, for maps scaled to show a whole state or a continent can indicate only generalities; but large scale soil maps, like large scale geological maps, are difficult to understand because they provide the local names for their soil types. One can expect to encounter in a local area Great Soil Groups not listed on a small scale soil map.

It is advisable to be well-prepared before going into the field. Both a geological and a soil map should be consulted. They can help with rock and soil identification respectively. If larger scale maps are unavailable, maps of Australia or of a single state will still be useful. A hammer should be taken (preferably a geological one) for breaking rocks and hard layers (such as laterite) in soils, a ruler (for measuring the depth of the horizons), pencils, and profile data sheets supported by a hard back for writing in the field. Use of a soil testing kit[1] enables a lot more information to be collected. The use of chemicals commonly found in soil testing kits will be explained in the appropriate sections.

Soil profiles can be suitably described on tabular data sheets. These ensure that no important property is omitted. To the pedo-

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[1] Soil testing kits issued to schools in New South Wales contain: a bottle of hydrochloric acid for testing for lime; Raupach Indicator and barium sulphate for measuring soil reaction (acidity); hydrogen peroxide for testing for organic matter and manganese; and silver nitrate, distilled water, funnel, test tube, and filter paper, for testing for chlorides.

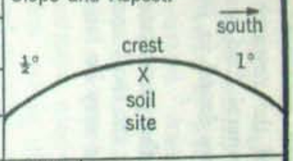
Location: *East Lakes Golf Course, Pagewood, Sydney*

Underlying Rock: *Wianamatta shale* | **Parent Material:** *dune sands*

Vegetation and Land-use: *Banksia shrubs, used as golf course* | **Drainage:** *good*

Soil Classification: *podzol* | **Date:** *1st June, 1968* | **Average Annual Rainfall:** *47.5″*

Slope and Aspect:



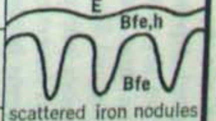
HORIZON	DEPTH	LOWER BOUNDARY	COLOUR	MOISTURE	TEXTURE	STRUCTURE	ORGANIC MATTER	pH	LIME AND SALTS	SPECIAL FEATURES
Ol	⅛″-0	gradational irregular	black	moist	—	platy	entirely decomposing leaves	4.0.	Nil	No mineral matter
Ah	0-10″	gradational even	dark grey	moist	sand	apedal single grain	strong root penetration	5.0	Nil	
E	10-50″	sharp irregular	white	dry	sand	apedal single grain	weaker root penetration	4.5	Nil	Sketch of horizons 
Bfe,h	50-54″ extending to 62″ in places	sharp very irregular	dark brown	moist	loamy sand	apedal massive	very weak root penetration	5.0	Nil	
Bfe	54-68″	gradational even	reddish yellow	dry	sand	apedal single grain	no roots	5.0	Nil	scattered iron nodules in **Bfe,h** and **Bfe**
C	68″+		light yellow	dry	sand	apedal single grain	no roots	5.5	Nil	dune bedding evident

Figure 2.2: Soil profile descriptive table.

logist, morphological properties are the most important; they are properties that can be described rather than measured. The properties selected depend on the purpose of the research, but the following are the most useful ones in general field studies. (More detail on these and other properties are found in Butler, 1955 and Brewer and Sleeman, 1960.)

Fig. 2.2 suggests one possible format for the presentation of profile data. This format may be varied to suit the local area; for drier areas columns headed *lime* and *chlorides* may be included.

SITE DETAILS

At the top of the tables are the details about the profile's site. These are very important as they can assist in the recognition of some of the soil-forming processes. Students should record the exact *location* of each site in case they wish to return to it or their teacher wants to check their work. Even where the site is not close to a road, a rough estimate should be made of its distance from the nearest road.

Both the soil's *parent material* and the *underlying rock* should be recorded. These will be the same only for soils developed *in situ*. The rock underlying deep soils on unconsolidated parent materials may not be exposed, so the student may leave this section unfilled rather than try to guess it. If he finds rock identification difficult, he should read through the section on rocks in his physical geography textbook, e.g., in Strahler, 1960. Unconsolidated (non-rock) parent materials may be grouped into four categories according to origin.

1. *Alluvium*. This includes all deposits laid down by streams. It is usually quite easy to recognise by its low position in the landscape (this does not always apply, e.g., in incised terraces and uplifted river gravels) and the presence of rounded water-worn pebbles. The texture of alluvium varies with distance from the stream, being coarsest nearer the stream. Sandy alluvium is usually found only in close proximity to the stream. The common widespread alluvial soils such as the grey and brown soils of heavy texture (see Chapter 10) are heavy or clayey in texture.

2. *Colluvium* (from the Latin *collis*, a hill). This refers to all materials carried downslope. Movement may be either slow (e.g., soil creep) or rapid (e.g., rock slide). Colluvial soils are usually deeper than neighbouring *in situ* soils and are usually aligned parallel to the slope. The bottom of the *C* horizon in such soils may contain a 'stone-line' which marks the limit of downslope movement. In all areas with slopes of more than 1 degree, colluvial soils should be looked for, especially if the vegetation is scanty.

3. *Aeolian materials*. These are various types of wind-transported materials. Vast areas of Eurasia and North America are covered by fine textured calcareous material called *loess*. In south-eastern Australia, the fine-grained calcareous wind-transported material is called parna. Both parna and loess blanket the landscape irrespective of slope or position, unlike alluvium or colluvium which are restricted by slope and position. In Australia, wind-transported soils are most widespread in the Riverina and in the desert; in the latter, both fine and coarse textured materials are to be found. Coarse textured saltated wind-transported materials are referred to as sand dunes or sand ridges, ridges usually being larger than dunes. Many coastal deposits are wind-transported, but since the amount of salt in them may be initially very high, they are best grouped with the *coastal deposits*.

4. *Coastal deposits*. These are generally sandy (they may contain gravels), but not necessarily quartzose. Many Australian beaches, especially those of southern Australia, contain appreciable quantities of calcareous material, while others, especially along basaltic coasts, may contain a lot of feldspar and heavy minerals. Quartzose sands are the most common marine deposits along the eastern coast of Australia.

All these unconsolidated parent materials, being pre-weathered, form deeper profiles than similar rock parent materials if given similar durations of weathering. Quite often the depth of the profile is the first indication that the parent material is unconsolidated. However, this generalisation should be used with caution, as some soils, such as lateritic soils and krasnozems, will be deep even on rock parent material.

When describing the *vegetation*, broad classes should be used, such as rain forest, wet sclerophyll forest, dry sclerophyll woodland, grassland or desert scrub, rather than actual species. Under *land-use*, there should be a note on whether the topsoil has been disturbed by ploughing, resulting in changes in the structure of the *A* horizons and possibly some mixing of the *A* and *B* horizons. Preferably, undisturbed soil profiles, such as are found in areas of unimproved grazing or forest, should be described.

Under the heading *drainage*, there should be a reference to the whole profile, that is, drainage of the site rather than internal drainage conditions within the profile relating to specific horizons (these are more appropriately noted in the column *special features*, within each horizon). Some soils contain hardpans causing the upper horizons to be poorly drained. The nature of the external drainage can be ascertained by the wetness of the profile several days after rain. Poorly drained soils remain wet, while well-drained soils dry out. The lower the position on the slope and the more impervious the substrata, the poorer are the external drainage conditions. A soil with a water table within the solum is classed as poorly drained. Two drainage categories, namely *free* and *poor*, may be sufficient.

The *soil classification* cannot be completed until the soil profile is described. Chapters 6 to 12 may be referred to for help in the classification of soil profiles.

The average annual *rainfall* of a local area may be obtained from the local Post Office. Students need to remember, however, that local rainfall can depart widely from the average and that there can be great variations over relatively short distances, so the average figure is only a rough guide to leaching effectiveness. Local Post Offices usually record daily rainfall and maximum and minimum temperatures. The figures for the week preceding a description of a soil profile will help the observer to study the drainage conditions and to evaluate the internal water movements through the profile.

In case weather data cannot be gathered before describing a soil, it is wise to record the *date*. Then one can check later to find whether the soil was described in, for example, a dry or wet season (average monthly rainfall figures can be obtained from the Post Office or from the *Commonwealth Yearbook*, which gives rainfall data for the main stations in Australia).

Slope and aspect can be recorded on a rough profile down the slope. In particular, any breaks in slope and the distance of the soil from these breaks should be noted. With practice, students should be able to estimate slopes to the nearest degree, which is all that is needed here. If this is not possible, students can make a rough theodolite from a plastic 180° protractor, and a ruler, with a plumb-bob made from a piece of string with a weight, such as a small sinker, on the end, as shown in Fig. 2.3.

Aspect refers to the direction the slope faces. This of course will not apply in level situations. There are sometimes quite marked differences in soil morphology arising from aspect differences. In the Southern Hemisphere, north-facing slopes are usually warmer, so weathering may be more intense. Students not in the habit of carrying a compass should determine aspect from the position of the sun. Only eight points of the compass (north, north-east, east, south-east, etc.) are needed.

Most of these site details relate to the variables which control the soil-forming processes. These variables are termed the *soil-*

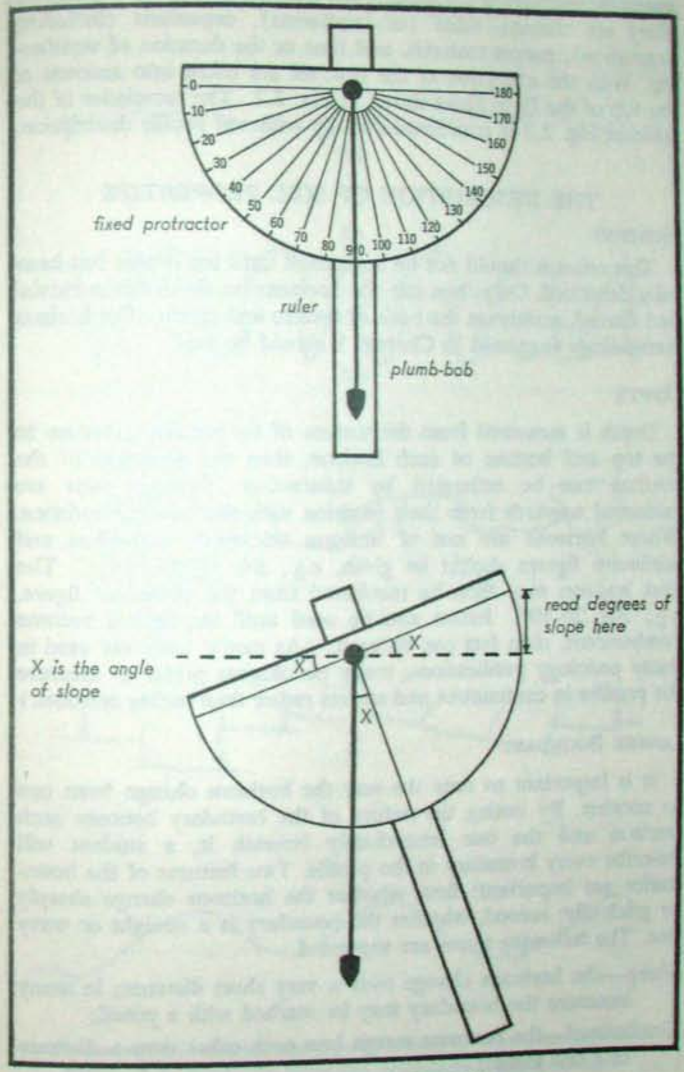


Figure 2.3: A simple theodolite.

forming factors and are discussed in Chapter 5. Five variables are generally considered responsible for soil morphological differences. They are climate, relief (or landforms), organisms (including vegetation), parent material, and time or the duration of weathering. With the exception of the last, all are taken into account at the top of the Data Sheet shown in Fig. 2.2. The remainder of the table in Fig. 2.2 is concerned directly with soil profile description.

THE DESCRIPTION OF SOIL PROPERTIES

HORIZON

This column should not be completed until the profile has been fully described. Only then can the horizons be divided into eluvial and illuvial, mainly on the basis of texture and colour. The horizon terminology suggested in Chapter 1 should be used.

DEPTH

Depth is measured from the surface of the mineral horizons to the top and bottom of each horizon, then the thickness of the horizon can be estimated by subtraction. Organic mats are measured upwards from their junction with the mineral horizons. Where horizons are not of uniform thickness, maximum and minimum figures should be given, e.g., *Bh* 11-20 (28)″. The next horizon may then be numbered from the minimum figure, e.g., *Bfe* 20-30″. Inches can be used until the figures become cumbersome, then feet can be used. (As metric units are used in many pedology publications, many pedologists prefer to measure the profiles in centimetres and metres rather than inches and feet.)

LOWER BOUNDARY

It is important to note the way the horizons change from one to another. By noting the nature of the boundary between each horizon and the one immediately beneath it, a student will describe every boundary in the profile. Two features of the boundaries are important: first, whether the horizons change sharply or gradually; second, whether the boundary is a straight or wavy line. The following terms are suggested.

Sharp—the horizons change over a very short distance; in many instances the boundary may be marked with a pencil;

Gradational—the horizons merge into each other over a distance of a few inches;

Even—the boundary between the horizons is parallel to the land surface;

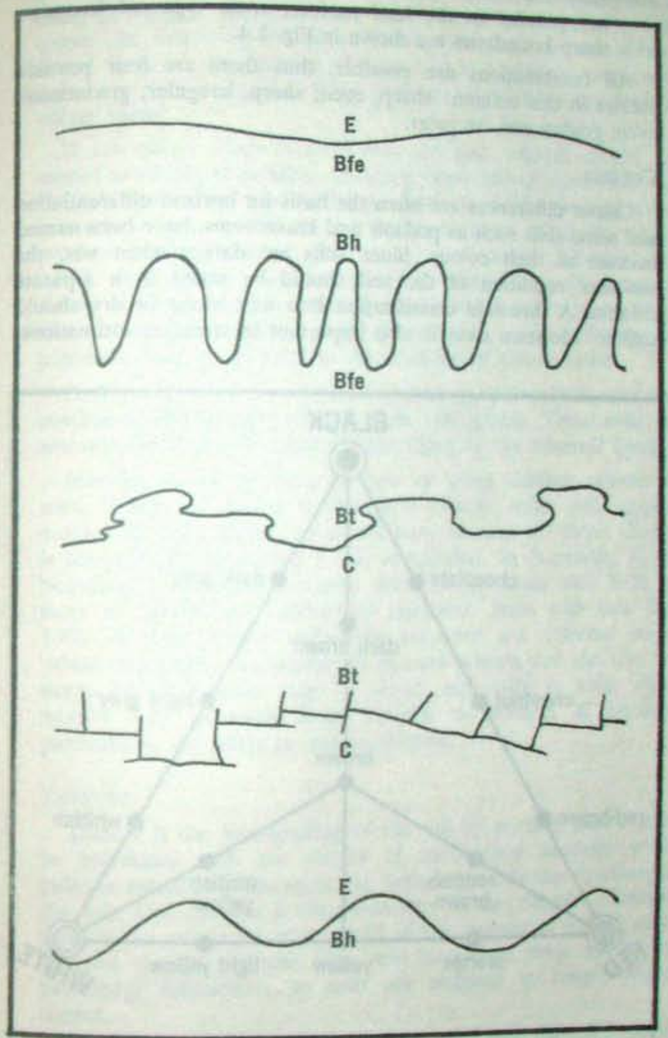


Figure 2.4: Types of irregular boundaries between soil horizons.

Irregular—the boundary between the horizons is neither straight nor parallel to the land surface; some types of irregular, sharp boundaries are shown in Fig. 2.4.

All combinations are possible, thus there are four possible entries in this column: sharp, even; sharp, irregular; gradational, even; gradational, irregular.

COLOUR

Colour differences are often the basis for horizon differentiation and some soils such as podzols and krasnozems, have been named because of their colour. Since soils are darker when wet, the *moisture* condition of the soil should be noted in a separate column. A threefold classification into wet, moist or dry should suffice. Moisture state is also important in structure estimations.

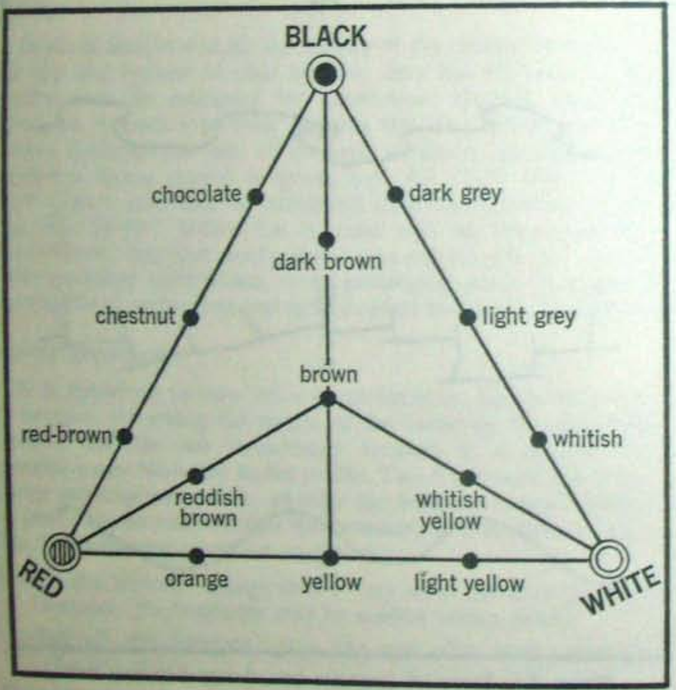


Figure 2.5: Graphic presentation of soil colour (well-drained soils).

The description of soil colour tends to be rather subjective, so many pedologists prefer to use a colour chart when assessing soil colour. In Australia, the Munsell Soil Colour Chart, an abbreviated form of the Munsell Colour Chart, is widely used, and the soil colour is given a value such as *5YR3/4*, rather than a colour name.

If no colour chart is available, the soil colours should be named as simply as possible, avoiding vague colour names such as "beige", "off-white" and "salmon pink". In freely drained soils, soil colour can usually be attributed to three pigments which produce black (humus), red (iron) and white (silicates and salts). Soil colours may be related to these three pigments as shown in Fig. 2.5 which presents ample colour names and clearly shows the relationship of each colour name to the three main pigments. This graph helps in the selection of colour names.

Poorly drained soils, in which reducing rather than oxidising conditions prevail, may contain blues and greens. These soils are accommodated on the Gley Colour Chart in the Munsell System.

Mottling is the presence of two or more distinct colours in soils. When soil colour is examined closely, most soils appear mottled to some degree, so an arbitrary amount of colour change is selected for designating a soil as mottled. In Australia, K. H. Northcote (1965) has reserved this term for soils with 10% or more of their matter differently coloured. Soils with less than 10% of their matter differently coloured are referred to as 'whole coloured'. The number of colours present and the size and shape of the mottles may be noted, especially in soils where mottles vary markedly from horizon to horizon. *C* horizons particularly, are likely to appear mottled.

TEXTURE

Texture is the field grading of the soil by particle size. It may be correlated with the results of mechanical analysis which indicate actual percentages of the various particle size fractions in the soil. This analysis is discussed later in this chapter. Texture, then, is the estimation in the field of the amount of gravel, sand, silt and clay in the soil. The field procedure does not permit percentage estimations, so soils are assigned to broad texture classes.

The *procedure* for estimating texture is to work the soil between the thumb and fingers until it is broken into individual particles, then to wet the soil to 'sticky point' (until it just adheres to the fingers) and note its properties after moulding. The texture classes relate to the degree of grittiness (indicating sand), silkiness (indicating silt) and moulding ability (indicating clay). Gravel is determined separately. The dimensions of each of these fractions are given later in this Chapter.

The texture term *light* is applied to soils with much sand and silt relative to clay, and *heavy* to soils with much clay relative to sand and silt. Sand has a greater ability than silt to make a soil light. The following texture classes grade from light to heavy and depend on the soil's ability to be rolled into rods about 3 inches long and ¼ inch across:

Sand	does not cohere	has a gritty feel	will not roll into rods — breaks easily
Loamy sand	cohesion just perceptible		
Sandy loam	coheres and is friable		
Silty loam	friable and coherent, with a silky feel		
Loam	friable and coherent; sand grains cannot be felt		
Sandy clay loam	has a gritty feel		rolls into a rod but cannot be turned into a ring without cracking
Silty clay loam	has a silky, smooth feel		
Clay loam	sand grains cannot be felt		
Sandy clay	has a gritty feel		
Silty clay	has a silky, smooth feel		
Light clay	easy to mould		rolls into a rod easily; can be turned into a ring without undue cracking; sand grains cannot be felt
Medium clay	fairly stiff and plastic to mould		
Heavy clay	very stiff and plastic to mould		

There is little point in estimating texture and structure for *C* horizons, so these properties apply to the solum only.

STRUCTURE

Soil structure may be defined as the arrangement of the soil or the form and orderliness of the soil mass. The essential element of soil structure is the *ped*. Peds are the stable, natural aggregates of soil structure, resulting from the soil's tendency to fracture along regularly spaced planes of weakness. When peds are the only aspect of soil structure described, it is better to use the term *pedality* rather than *structure*, of which peds are only one aspect. Other aspects of soil structure which may be considered are the nature of pore spaces or voids in the soil, the stability of the soil aggregates in water, soil porosity, and bulk density. Most of these are laboratory determinations, so the pedologist is concerned primarily with pedality.

Pedality may be demonstrated by throwing a shovelful of soil against a wall and noting the aggregates on the ground after the soil has fallen—these are peds. In field work, peds are most clearly observed in the dry soil of gully faces before the soil is disturbed. In soils exposed by augering, the peds are usually destroyed, so in describing augered soils, the 'structure' Column is usually left unfilled.

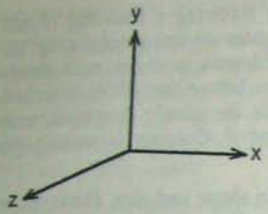
Peds are generally fairly uniform in shape and size. However, in many soils, there may be two or more distinct types of peds. These are called primary, secondary and tertiary peds. For example, the *B* horizon of a red-brown earth soil may contain primary peds, 9″ by 3″ by 3″ columns. These peds may break into 1″ blocks, the secondary peds, which may further fracture into tertiary nutty peds. In this soil, all three levels of structure should be described, but in many soils only one level of peds may be apparent.

Soils which contain peds are referred to as *pedal*, and so, conversely, soils which lack peds are referred to as *apedal*. Apedal is a more useful term than 'structureless' which implies no arrangement at all. Apedal soils have two distinct arrangements. One is *apedal, single grain* which is found in all dry sands. The opposite extreme is *apedal, massive* which is found in wet, clayey soils which coalesce completely. In apedal, massive soils, there are insufficient voids to cause separation into peds, so in clayey soils, pedality usually cannot be determined if the soil is very wet. *Fragments* are soil pieces which have no uniform shape or size and so cannot be termed peds. *Clods* are soil aggregates caused by ploughing or other agricultural disturbances. They are large lumps of soil which tend to peel when dry.

Pedality may be estimated by describing the frequency, size and shape of the peds, and their breakdown into secondary peds. The *frequency of the peds* is often referred to as *grade* of structure. It is useful to assess how much of the soil is in peds and how much is apedal or fragmental. This may be done on a rough percentage basis (e.g., 30% blocky).

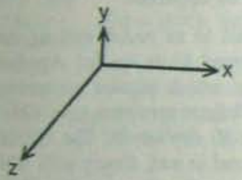
The *names of the peds* vary from country to country so a mathematical terminology is most useful. Some of the smaller ped names include sizes, but the size of larger peds should always be specified, in inches or centimetres. Peds may be divided according to the nature of their *three* axes into:

1. Cubical — *x*, *y* and *z* axes roughly equal.

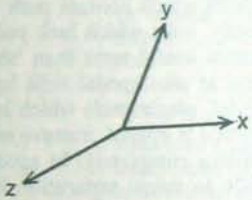


2. Prismatic — *y* axis long, *x* and *z* axes shorter and roughly equal.

3. Laminar — *y* axis short, *x* and *z* axes longer and roughly equal.

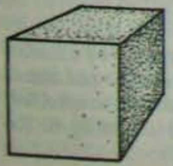


4. Parallelipipedal — the *x*, *y* and *z* axes are not all at right angles. May be skewed in one or two directions.

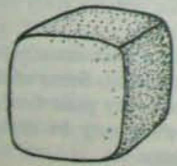


1. *Cubical.* There are two main groups in the cubical system, depending on whether there are six or more faces. Spheroidal structure, which has no faces, is the end-member of each system, as shown:

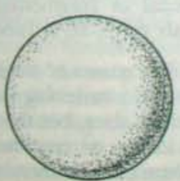
Blocky, also called cubic, bounded by six faces intersecting at right angles.



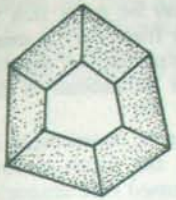
Sub-angular blocky, bounded by six faces intersecting with rounded interfaces.



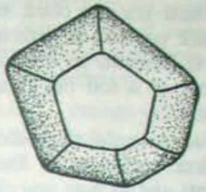
Spheroidal, no planar faces.



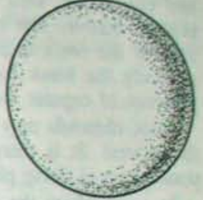
Polyhedral, more than six faces with sharp planar interfaces.



Sub-angular polyhedral, more than six faces with rounded interfaces.



Spheroidal, no planar faces.



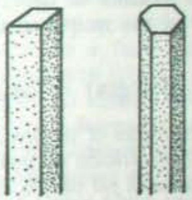
Nutty—fairly angular, polyhedral peds, ½"-¼" diameter.

Granular—rounded peds, less than ¼" diameter, fairly porous and stable.

Crumb — rounded peds, less than ¼" diameter, unstable and crumbling into smaller units.

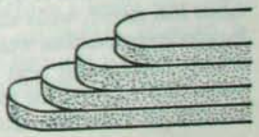
2. *Prismatic.* There are two main groups in this system, depending on whether the interfaces are angular or rounded:

Prismatic, flat tops and bottoms to the columns, angular sides. May be an extension of the blocky (four sided top) or polyhedral (five or six sided top).

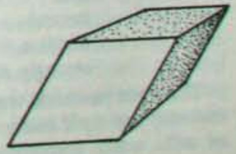


Columnar, rounded tops and interfaces on the sides.

3. *Laminar,* also called platy or squamose. The peds are found in sheets. There is only one definite plane of weakness.



4. *Parallelipipedal.* The name refers to the shape of the solid, skewed parallelogram. The interfaces are generally angular. The most common skewed ped is in the blocky form, the parallelipipeds often being several inches across.



ORGANIC MATTER

The black pigment in soils is usually derived from organic matter; however, black soils are not always rich in organic matter, and conversely, soils which are not black may be quite rich in organic matter. The black pigment is derived from only a small part of the soil's organic fraction, so it is important to observe not only the black colour of a soil but other indications of the presence of organic matter.

Root channels may influence the deposition of constituents in the subsoil. It is therefore important to find the depth of root penetration of living plants and to note any disused root channels.

In cases where the activities of soil fauna such as earthworms have a noticeable effect on soil morphology, their presence should be noted in the column headed 'organic matter'.

Black humus in soils effervesces when hydrogen peroxide is applied. The amount of effervescence gives a rough indication of how much humus is present.

REACTION or *pH*

Soil reaction is the term used to describe soil acidity or alkalinity. Since soils are only mildly acid or alkaline, the *pH* scale is the most suitable measure of their reaction. *pH* is the logarithm to the base 10 of the reciprocal of the hydrogen ion concentration, or

$$pH \text{ (potens hydrogen)} = \log_{10} \frac{1}{[H^+]}$$

Thus *pH* is the negative logarithm of the hydrogen ion concentration. Since water has a concentration of hydrogen ions equal to 10^{-7} moles per litre, water has a *pH* of 7.0. Acid substances have higher concentrations of hydrogen ions than water and so have *pH* values less than 7, while alkaline substances have lower hydrogen ion concentrations than water and so have *pH* values more than 7.

Soils, being mildly acid or alkaline, generally lie between *pH* values of 4 and 10. The following table indicates the relationship between *pH* value and reaction for soils:

strongly acid	*pH* up to 5.5
acid	*pH* 5.5 to 6.5
neutral	*pH* 6.5 to 7.5
alkaline	*pH* 7.5 to 8.5
strongly alkaline	*pH* above 8.5

There are two main ways of measuring soil *pH*. In the laboratory one may use a *pH* meter with glass electrodes and measure the *pH* of soil: water suspensions in certain ratios (e.g., 1:5 or 2:5).

Portable battery-operated field *pH* meters are available, but so far in Australia they are not widely used. In the field, *pH* is usually measured with indicators which give much less accurate values than *pH* meters. Even Raupach's Indicator, used in schools' Soil Testing Kits, gives differences only of 0.5 *pH* units. Raupach's Indicator is the most suitable with soils as there are marked colour changes over the *pH* range 4 to 9 as shown:

pH 4.0	orange-yellow
pH 4.5	yellow
pH 5.0	greenish-yellow
pH 5.5	yellowish-green
pH 6.0	light green
pH 6.5	mid green
pH 7.0	greyish-green
pH 7.5	greenish-grey
pH 8.0	purplish-grey
pH 8.5	greyish-purple
pH 9.0	purple

Raupach's Indicator is a mixture of three indicators (brom-cresol green, brom-cresol purple and phenol red) which change colour over the *pH* range listed above. The *procedure* is to place about 1 gm of soil onto a clean spotting tile (a saucer will do), add enough Indicator to make a thick paste, and mix thoroughly. Sprinkle the paste with barium sulphate powder (which must be at exactly *pH* 7.0), and after 3 minutes read the colour of the barium sulphate with the aid of a Raupach Colour Chart. The colour is read from the barium sulphate powder, rather than the soil, whose own colour may mask the colour of the Indicator.

The next two soil features (lime and chlorides) are not listed in Fig. 2.2 because the soil described is from a humid area and is therefore thoroughly leached of these constituents. However, in soils of arid and semi-arid areas, both lime and chlorides should be looked for, and columns should be included on the soil data sheet for their description.

LIME

Lime refers to calcium carbonate ($CaCO_3$) which is present in most soils of the drier regions. Its presence is shown by the application of 10% hydrochloric acid to the soil. Lime effervesces immediately with hydrochloric acid. The acid test should be used on all soils of drier areas, as many will be found to effervesce even though no white flecks of lime are visible. Lime may be present in soils in the following forms:

1. As *nodules*, which are white lumps of calcium carbonate of variable size. Nodules are found in lenses in certain horizons or may be scattered fairly uniformly throughout an horizon. The average size of the nodules should be specified.
2. As *fine earth*, which may or may not be visible. Some soils are so heavily charged with lime that the fine earth separates as white powdery calcium carbonate. In other soils, this powdery lime may be finely disseminated and can be detected only after the application of acid.
3. As *floaters* of rock in soils derived from limestone. These floaters, like lime nodules and fine earth, effervesce with acid.

CHLORIDES

In arid areas salts accumulate in soil profiles. One of the most common is sodium chloride or common salt which may not always be visible as white specks. Test for chlorides by placing about 4 gms of soil in a filter paper in a funnel over a test tube. Wash the soil with distilled water and then add silver nitrate to the filtrate in the test tube. If chlorides are present in the soil, a white precipitate will form in the filtrate. The thickness of the precipitate is a rough guide to the amount of chloride in the soil.

SPECIAL FEATURES

There are several properties which should be noted in this column, but which are not listed separately because of their rather restricted occurrence.

Stones in the soil should be described, noting the rock type, size, shape, frequency and whether they are aligned in any particular direction.

Nodules not so far described, such as sequioxide nodules, should be noted, especially their size, colour and frequency. *Concretions* are nodules which appear concentric about a nucleus: a student can usually identify them readily by breaking open some nodules and examining them.

Clay Skins, the shiny surfaces on wet peds, should be noted. Their colour should be described (relating it to the host soil), and some idea of thickness should be given. (In case the soil is not wet, or is not near a creek, a student should carry a bottle of water.) Black humus and manganese skins effervesce when hydrogen peroxide is applied.

Any other point of interest not covered by the other columns should be mentioned in the "Special Features" column.

THE SIGNIFICANCE OF SOIL PROPERTIES IN RELATION TO THE SOIL-FORMING PROCESSES

Variations in the soil-forming processes, and hence in the soil properties, depend on the five factors of soil formation (climate, relief, organisms, parent material and time). A student's profile descriptions should take into account these five factors by describing neighbouring profiles. He might describe a series of soils along a slope, starting with a shallow stony soil on the upper slope and moving through colluvial soils until, possibly, the *in situ* soils on the gentler lower slope are reached. Then he could note the changes in the properties of the soils down the slope. The following material is intended to apply to several profiles in the one area, rather than to a single profile description. Profile description will be more valuable if many profiles are examined, showing variations within a single area of the various soil-forming factors.

DEPTH

Generally, the steeper the slope, the shallower are the soil horizons. A useful exercise is the graphing of the depth of a certain horizon, such as the *E* horizon of a podzolic soil, against the angle of slope.

Pre-weathered parent materials produce deeper profiles than chemically similar rock materials. The greater the rainfall, generally, the greater the depth of weathering.

LOWER BOUNDARY

Very *sharp* boundaries between horizons often signify marked changes in all properties of the soil. This is true especially in the *E/B* boundary of podzolised soils where the colour, texture, structure, organic matter and *pH* all change abruptly. Gradational boundaries are usually associated with less abrupt and less intense changes in the soil properties. Over some gradational boundaries, there may be no colour change and only slight changes in texture, structure and *pH*. A gradational boundary often provides insufficient change in the soil environment for the mass deposition of eluviated constituents.

Irregular boundaries often follow root channels, indicating a relationship between illuviation and vegetation. Some irregular boundaries may be related to heterogeneous parent materials.

COLOUR

Soil colour, the most obvious property and the one easiest to describe is perhaps the most difficult to interpret as there is much controversy over the exact chemical composition of the main colour pigments. To say that black colours are due to humus is imprecise, as only a small part of the soil's humic fraction serves as a black pigment. It is thought that the *humic acid* fraction of organic matter produces the black colour in soils. Other fractions, such as fulvic acid or insoluble humin, are only slightly coloured. Thus soils such as the chernozems, which may not contain a large amount of total organic matter, are black in colour because much of the organic matter exists as humic acid. Other soils in which the organic matter is not well humified, may contain large amounts of total organic matter and yet not be black at all.

An additional problem is the masking of colour pigments. Some soils may contain so much humic acid that any other colour pigment, such as iron oxide, may be masked. Other soils may contain much less iron oxide than these soils and yet be quite red in colour. Thus very little can be learned about soil colour from total organic matter or total iron oxides. The form of these is far more important than their quantity.

The effects of iron as a soil colour pigment are not as simple as many textbooks suggest. Iron oxides tend to form amorphous, colloidal coatings on the soil grains, and, as the percentage increases, crystalline forms may develop. In the laboratory, a series of *iron oxides* exists, from the unhydrated member, haematite (Fe_2O_3), to the highly hydrated member, limnite ($Fe_2O_3 \cdot 3H_2O$), with intermediate members such as limonite ($Fe_2O_3 \cdot 3/2H_2O$). It was formerly believed that red colours arose from the unhydrated iron oxide, haematite, and yellow colours from the more hydrated members, but P. E. Raymond showed in 1927 that soil reds are derived from less hydrated members (4% to 6% water) and yellows from more hydrated members (6% to 30% water), so it seems that truly unhydrated members in soils are rare.

The red and yellow hydrated iron oxides are weathering products of any mineral containing iron, e.g., olivine, as well as of the primary iron minerals such as magnetite. Also, most sedimentary rocks contain some colloidal hydrated iron oxides. Of the soils on igneous rocks, it may be assumed that yellow soils form under wetter conditions than neighbouring red soils. For this reason, red soils tend to form on well-drained slopes, while yellow soils tend to develop in poorly drained depressions. However, the colours may be reversed in soils formed from sedimentary rocks: sometimes upper slope soils will be yellow because they are derived from rocks high in yellow iron oxides, while lower slope soils may be red because their parent material is higher in red iron oxides. Thus the matter of inheritance of iron oxides from the parent material needs to be considered as well as the formation of the soil *in situ*.

In profiles where iron oxides are being illuviated, redder colours usually prevail in the upper *B* horizon while a tendency to yellow is often found in the lower *B* horizon, indicating wetter conditions at that level. Many lower *B* horizons of red soils are also found to be mottled with yellow, probably indicating periodic invasions of the lower solum by the water table, followed by drying.

Reduction and oxidation as well as the degree of hydration are important in producing colour pigments. Prolonged invasion by the water table produces drab greys and yellows and sometimes even blues and green from the ferrous (iron II) oxide (the oxides producing the red and yellow colours are of ferric or iron III oxide). Saturation of the soil for long periods can affect any horizon. The consequent bleached effect is called *gleying* and the resultant horizons are known as gley horizons. Many of the iron oxides have migrated from the gley horizons, and those that remain are pale in colour. Gley horizons feature in many soils of the ill-drained, glaciated areas of North America and Europe but they are not very common in Australia. A feature similar to gleying, but more intense, is the formation of the pallid horizons in a laterite profile: a marked bleaching occurs and such horizons are thought to have developed permanently below the water table.

It is wrong to relate all black colours in soils to humic acid. Many other common soil constituents are black, especially manganese oxide (MnO_2). Manganese oxide clay skins are common in some soils, but are usually more lustrous than organic matter skins. Purely manganese oxides and hydroxides are rare in soils, so many of these coatings are probably complexes of manganese with other substances such as aluminium, potassium, cobalt, magnesium and calcium.

There are many substances that tend to produce white colours in soils. Most pure clays are white; where red appears, there is probably an iron ingredient. The aluminium oxides, which form a series similar to the iron oxides, are also white (the reddish colour of the ore, bauxite, which is mined for aluminium oxide, derives from its iron content not its aluminium). Lime ($CaCO_3$), gypsum ($CaSO_4 \cdot 2H_2O$), and unstained quartz are also white.

TEXTURE

Field determinations of texture give only rough estimates of particle size grading in soils. The ambiguous term *mechanical analysis* refers to the more accurate laboratory assessment of particle size grading with which texture may be compared.

There are many systems of particle size grading currently in use, but the tendency is towards the International (or Atterberg) System which has very simple divisions. The names gravel, sand, silt and clay are size gradings and any differences in minerals between these gradings is coincidental. The International (Atterberg) fractions are as follows:

	mm diameter
Gravel	more than 2.0
Coarse sand	2.0—0.2
Fine sand	0.2—0.02
Silt	0.02—0.002
Clay	less than 0.002

Investigations show that not many primary (unweathered) minerals are present in the clay fraction, whereas the sand fractions are mainly primary minerals. However, aggregates of clay minerals, especially the iron oxides, may remain in the sand fraction during mechanical analysis. Clays are disperse systems of the colloidal products of weathering in which secondary mineral particles of dimensions smaller than 2 microns predominate.

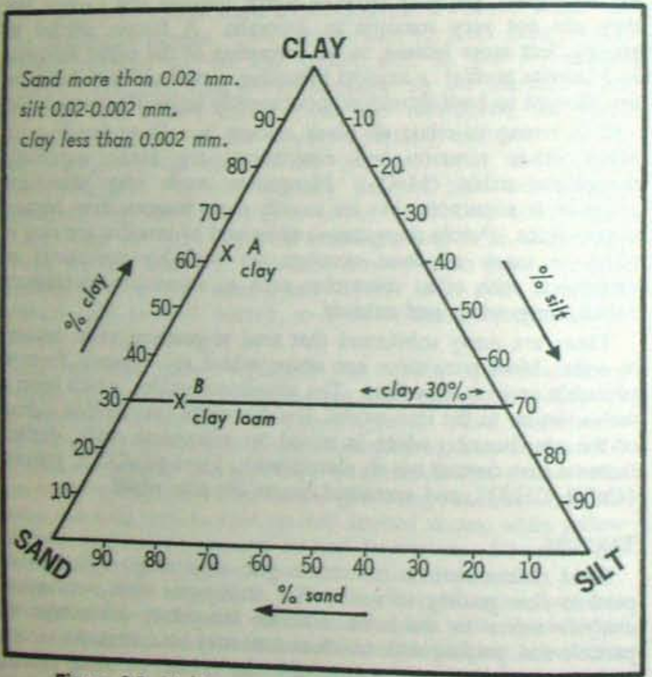


Figure 2.6: Plotting mechanical analysis results on the texture triangle.

In the Appendix, a quick method for the mechanical analysis of soils is given. The results of such analysis are given as percentages of the various fractions, e.g.,

	Gravel	Coarse Sand	Fine Sand	Silt	Clay	Total	Field Texture
Soil A	nil	15.1	17.9	5.4	62.0	100.4	clay
Soil B	3.0	32.5	24.1	10.3	30.0	99.9	clay loam

There should be quite close agreement between the field determination of texture and the mechanical analysis. The analysis results may be plotted graphically on a *texture triangle*, the vertices being sand (add the gravel, and coarse and fine sand fractions), silt and clay. The figures are always plotted parallel to the side opposite the constituent so that the line representing 30% clay in Fig. 2.6 joins the outside points 30% clay to 70% silt. In Fig. 2.6, soils A and B are plotted, and their field textures are

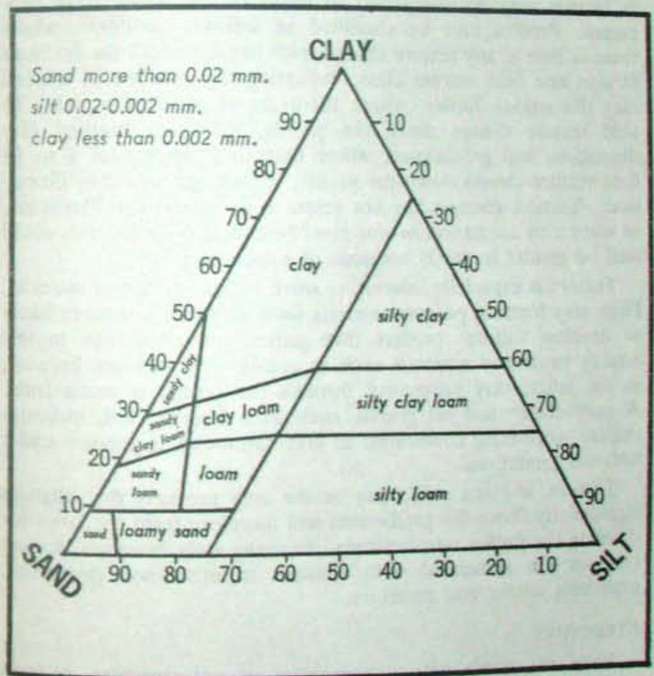


Figure 2.7: Marshall's diagram relating mechanical analysis percentages to the field determinations of texture (Australian soils).

also listed. After enough soils have been textured and mechanically analysed, all the results may be plotted to draw a *texture diagram* which indicates limiting percentages for each texture class.

Fig. 2.7 is derived from Marshall's 1947 diagram which showed the result of his analysis of Australian soils. It differs slightly from similar overseas diagrams because it gives less prominence to the silt fraction (which is less important in Australian soils). This diagram can be used to find rough percentages for your field textures. It omits some texture classes, and further subdivision into light, medium and heavy clays may be useful.

Texture gives a great deal of information about soil profile development, so it features strongly in many soil classifications—the primary divisions of Northcote's (1965) Australian soil classification are based on texture changes down the profile. Changes in texture may be interpreted in terms of the soil-forming processes. Profiles may be classified as follows: *uniform*, where there is little if any texture change with depth and all the horizons fit into one field texture class, indicating the absence of marked clay illuviation; *duplex*, where there are changes of at least 1½ field texture classes down the profile, indicating marked clay illuviation; and *gradational*, where there are changes of 1 to 1½ field texture classes down the profile, indicating some clay illuviation. Texture changes do not relate entirely to clay illuviation, as some clay formation *in situ* must be considered, and this could well be greater in the *B* horizons of some soils.

Texture is especially interesting when related to parent material. High clay-forming parent materials such as basalt are more likely to develop uniform profiles than parent materials high in less readily weathered minerals such as granite or sandstone because, in the latter, clay movement through the profile is much freer. A very clayey soil on granite such as a lateritic soil, indicates intense weathering conditions, as clay formation is greatest under hot, wet conditions.

Texture, in some soils, may be the only property that changes significantly down the profile and will therefore form the basis for dividing the profile into horizons. In many soils, however, texture changes are associated with changes in other soil properties, especially colour and structure.

STRUCTURE

Peds, the stable natural aggregates of soil structure, re-form when the soil dries out after a wetting. There are two steps in the formation of peds: flocculation and cementation. *Flocculation* is a temporary, electro-kinetic effect whereby the soil becomes aggregated, and if the flocculating agent is removed, the soil soon becomes dispersed (the reverse of flocculated). Flocculation, then, is only the starting point of ped formation, but it is an essential starting point, for without it the peds will not form.

It has long been recognised that lime has a beneficial effect on the soil's physical constitution as well as on its nutrient status. Calcium-saturated soils are more flocculated than sodium-saturated soils, so calcium may be regarded as a flocculating agent and sodium as a dispersing agent. The effects of these cations is not a simple one, as an excess of either may produce reverse effects. A soil such as a solonchak, which is saturated with sodium and contains, as well, excess free sodium, is flocculated rather than dispersed. However, removal of the excess sodium, leaving the colloid saturated with sodium, results in dispersion, as in the solonetz soils.

The addition of *lime* to a soil (especially to replace sodium) will improve the soil's structure only temporarily, for it is often quickly leached out and degeneration then occurs. Today calcium is not esteemed as highly for its contribution to soil granulation as it was in the past: soil scientists have shown that when calcium is replaced by sodium in soils containing enough humus there is usually no deterioration in granulation. However, the replacement of calcium by sodium in soils low in humus does cause marked structure deterioration. Thus flocculation is only the starting point of soil structure.

The second step in ped formation is *cementation*, whereby the unstable floccules are cemented into stable peds. There are several cementing agents for peds. The three most important are the silicate clay minerals, the organic colloids, and hydrated oxides of iron and aluminium; soils rich in all three are likely to contain stable peds. Pure sands, which lack all three, are apedal, single grain. Farmers have realised for a long time that the addition of organic matter has a beneficial effect on granulation in soil, but the beneficial effect diminishes soon after the addition, sometimes leaving the soil even worse off than before, unless the organic matter is replenished at regular intervals. *Iron oxides* exert a double effect on soil structure: they are both flocculating and cementing agents, because free iron in solution causes flocculation and, once dehydration takes place, the insoluble iron oxides that are then formed are cementing agents. The granulating effect on iron oxides is noted especially in tropical soils such as krasnozems, which, although poorly off chemically, are among the best-structured soils in the world.

The binding effect of *humus* is seen in the high degree of granulation in chernozems which compare with krasnozems in

CHLORIDES

The presence of chlorides in a soil also indicates incomplete leaching—in fact, much less leaching than in soils in which lime is the only salt. The first indication of chlorides may be a high *pH* value, e.g., 9.5; however, soils in which there are appreciable quantities of neutral salts, such as sodium chloride, may have a *pH* closer to 7. Therefore one must consider the possible accumulation of chlorides even if the *pH* is not high.

Chlorides are often encountered in the *B* horizons of arid soils where they are illuviated, but since they are readily soluble (unlike lime) they may be found anywhere in the profile, especially at or near the surface where they may be deposited as the soil water evaporates.

High concentrations of chlorides are often associated with poor vegetation or salt-tolerant plants such as salt bush or blue bush. Since chlorides are soluble, they may move quite rapidly through the soil profile, being deeper after rain, and nearer the surface in dry periods when they rise by capillarity and when plants return them to the surface as litter. Where micro-relief, such as gilgai, is present, chlorides often concentrate in the depressions.

SPECIAL FEATURES

Not all nodules and concretions in soils can be related to the current period of soil formation. This applies especially to the hydrated oxides of iron, as once these develop nodular or concretionary form, they may persist through several periods of soil formation. This is particularly true of nodules and concretions from laterite. These may take part in several consolidations before they are eventually destroyed, by weathering or transport.

However, there are usually morphological differences between primary and secondary nodules, the latter often being mixed with extraneous material.

As nodules weather in the soil they may give rise to mottles which are quite unrelated to water table fluctuations. Mottles formed in this way can usually be detected quite easily and often the centre of the mottle is harder where the nodule is less weathered.

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Chapter Three

Constituents of Soil

CONTENTS

SOIL IS COMPOSED OF FOUR CONSTITUENTS: MINERAL MATTER, organic matter, water and air. The division of soil constituents into these four categories is not a simple matter, as there is considerable over-lap. For example, dissolved in soil water are minerals existing as ions (charged bodies), and some of the products of the decomposition of organic matter is also ionic. Chemical weathering of soil mineral matter depends on the presence of water. The growth of vegetation, and hence the production of organic matter, depends on both water and soil air. Clearly, the four constituents are closely related, but for simplicity they are considered separately.

SOIL AIR

Baver (1966) considers that restricted aeration is the greatest limiting factor in the development of an extensive root system. Restricted aeration impairs root respiration which retards water and nutrient absorption and impairs the orderly functioning of the essential biological processes associated with good fertility.

Soil *air capacity*, which may be defined as the quantity of air (usually expressed as a percentage of the total soil constituents) remaining in the soil at field capacity, has been found to correlate with plant yields. There is an optimum air capacity for plant growth. Sandy soils, which contain large pores, have too much aeration at the expense of an adequate water supply; on the other hand, clayey soils have low air but high water capacity, and this is why farmers must continually struggle to aerate such soils by means of well-chosen cultivation methods. However, a clayey soil that is well-structured does not have very serious aeration problems. Air capacity is related to the size of the peds. For every soil, there is an optimum ped size for maximum air capacity, and the formation of larger and smaller peds decreases the air capacity.

Tilling or the addition of organic matter, cinders or sand can increase the air capacity of compact soils, while trampling and rolling can decrease soil air capacity. Trampling between rows in glass-houses has been found to decrease soil air capacity by 16% to 20%, causing serious problems.

Not a great deal is known about the air capacity requirements of plants but we know that plants vary in their tolerance of low air capacities. Sugar beet and potatoes require high air capacities, probably between 15% and 20%. Wheat and oats probably need at least between 10% and 15% air capacity, but their yields are greatly improved when the air capacity is raised to more than 30%. Relatively small increases in soil air capacity have been found to greatly increase the yield of corn in the United States.

The *composition* of soil air is no less important than the quantity of air in a soil. The soil atmosphere differs from the free atmosphere mainly in the percentage of carbon dioxide. Plant roots and micro-organisms living in the soil remove oxygen and respire carbon dioxide, thus the soil air is richer in carbon dioxide and poorer in oxygen than the free atmosphere. The extent of this difference depends on the rate of the biological processes and also on the rate of renewal of the soil air by atmospheric air. The mean composition of soil air, compared with atmospheric air is roughly as follows:

	Percentage in Soil Air	Percentage in Atmospheric Air
Nitrogen	79.2	79.0
Oxygen	20.55	20.97
Carbon dioxide	0.25	0.03

These values for the soil vary enormously from season to season; they also vary with the type of soil, and the kind of crop. The soil properties which influence the soil's air capacity (texture, structure, water and organic matter) also influence the composition of the soil air. If crops and vegetation remove oxygen more rapidly than it can be replaced, the composition of the soil air deteriorates. In waterlogged, poorly structured, heavy textured soils, the rate of renewal of soil air is extremely low. Under these conditions, which are termed *anaerobic*, there is more carbon dioxide relative to oxygen than is healthy for normal plant growth. In some anaerobic soils, there may be as little as 1% oxygen and more than 15% carbon dioxide.

Continuous cropping (i.e., no fallowing) increases the amount of carbon dioxide in soils. Soil air from continuously cropped land may contain ten times as much carbon dioxide as similar fallow land. Biological activity increases the carbon dioxide content; in fact, carbon dioxide content is used by soil microbiologists as a measure of biological activity. There are marked seasonal fluctuations, even in uncropped land, in the carbon dioxide content of soil air. These fluctuations are related mainly to temperature and moisture changes. When soils dry out in summer, the oxygen content of the subsoils increases substantially.

The composition of soil air changes mainly by the renewal of soil air with atmospheric air, as most photosynthesis occurs above the soil. According to the kinetic theory of gases, gas molecules are in such a state of agitation that two gases will readily mix as the molecules of one move into the space occupied by the other. This process is termed *diffusion*. Since soil air contains more carbon dioxide and less oxygen than atmospheric air, there is a strong natural tendency for the carbon dioxide to diffuse into the atmosphere and the oxygen to diffuse back into the soil air.

Diffusion is related more closely to soil factors than to meteorological factors. Its rate is reduced to one quarter as the free pore space in the soil is reduced to one half. Pore space is the dominant variable affecting the rate of diffusion which operates continuously even in heavy clayey soils, as long as the soil's pore space is not reduced to zero.

SOIL WATER

Water is essential for all forms of life. It might even be argued that, since plants can grow in nutrient solutions without soil (hydroponics), water is more important than soil. All soils contain some water, even in the driest desert, but it may be held so tenaciously by the soil that plants are unable to extract it.

Water is held in soils in two ways, firstly in the interstices (the pores or holes) between the soil particles, secondly by adsorption[1] onto the surface of the soil's clay particles. It is very difficult to separate these two types of water, but it is generally agreed that the first type is more readily removed from the soil than the second.

Water moves in all directions in soil. Wherever there is a hydraulic head, it will respond to gravity along the easiest possible route, which need not be downwards vertically. However, water can also move through soil against the pull of gravity as many of the pores in soil are so small that they serve as *capillaries* through

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[1] In soil science, the word 'adsorbed' is not to be confused with 'absorbed'. Absorbed means 'taken into' in the way in which calcium chloride takes in water from the atmosphere. Adsorbed means 'attracted' in the way iron filings are attracted to a magnet. 'Adheres' is close in meaning to 'adsorbs'.

which water is transmitted by surface tension. To demonstrate this, a student can place a piece of glass capillary tubing into a beaker of water. In a very short time he will notice that the water has risen in the tube (see Fig. 3.1).

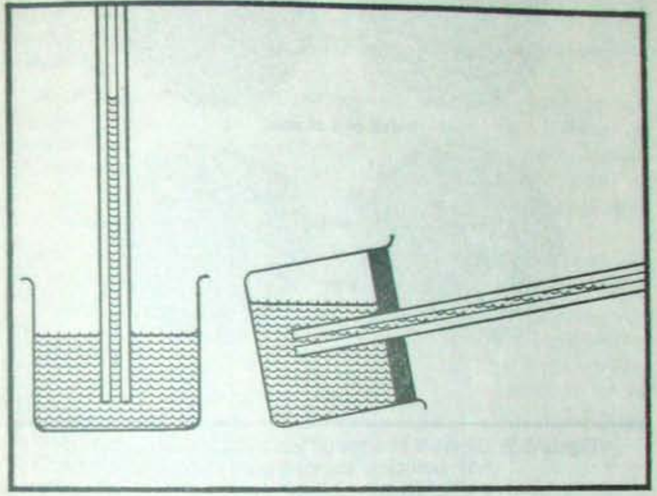

Figure 3.1: Water movement in glass capillary tubes.

If the tube is inserted into a stoppered vertical or horizontal vessel, it will be possible also to note that capillarity operates in all directions. Water moves through soil in a similar manner, and capillarity is certainly not restricted to upward vertical movements, though these do occur. However, unlike a glass capillary tube, the capillaries of soil are not always neatly interconnecting and are certainly not straight. The soil pores also contain air, which remains even though the soil has been thoroughly wetted. Capillary water exists in soils as thin films or sheets, very irregular in shape and thickness, sometimes bounded by solid particles and sometimes by air.

When a soil is wetted from above by rain or spray irrigation some of the air in the pores is replaced by water, and the rate of drainage of the soil depends on the diameter of the narrowest part of the soil pores. When a soil is drying out or is wetted from beneath (channel irrigation or a rising water table), the rate of rise of water depends on the size of the widest part of the soil pores, as this controls the rate of rise by capillarity.

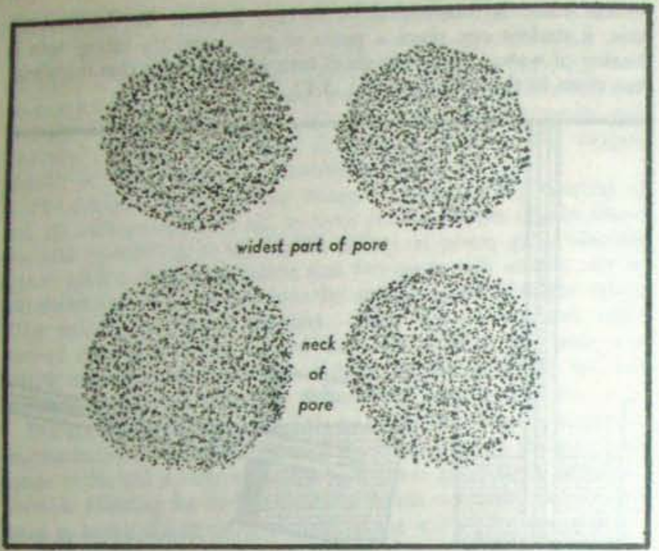

Figure 3.2: Diagram of a simplified soil pore, bounded by four particles, showing pore neck and widest part of pore.

A great deal is learned about the water in soil by measuring the force needed to remove it. After rain falls and the soil becomes saturated, the surplus water is very loosely held and soon drains. As the soil dries out, the rate at which water leaves the soil gradually diminishes, until downward movement (drainage) ceases and upward movement (capillarity) is the main way by which water is lost. At length a stage is reached where very little water can be lost, for it is held so firmly in the soil. At this stage, the soil becomes too dry to support plant life. Clearly, water is held at various pressures in the soil. Superfluous water is easy to remove because it is held at a very low pressure, but when considerable drying has taken place the remaining water is held at such strong pressure or tension that plants are unable to free it for their use. Nevertheless, even at the latter stage, considerable water remains in the soil.

Water loosely held in the soil may be measured in centimetres of pressure. By subjecting wet soil to certain pressures, one can measure the force necessary to extract some of the water. The apparatus for this experiment involves a water-filled burette which

can be lowered to apply a suction to the soil, so that the pressure can be measured as the *height of a column of water in centimetres.* Since there is a limit to the manageable height of a column of water, this measurement applies only to water quite loosely held in the soil. Water which is held at intermediate tensions in the soil is more readily measured in *atmospheres* of pressure. A 1,000 centimetre column of water approximately equals 1 atmosphere pressure.

A more useful scale for measuring the pressure at which soil water is held is a *pF* scale. This scale does not become as cumbersome for high pressures as either of the above scales. *pF*, like *pH*, is a scale designed for simplicity of measurement.

$pF = log_{10}$ of the height of a column of water in centimetres
p means 'power' and *F* means 'freedom'.

Thus a *pF* of 3 implies that the water is under a suction equal to the pressure exerted by a column of water 1,000 centimetres high.

The Appendix of this book provides some experiments on soil water. These are concerned with finding the amount of moisture remaining in a soil at various *pF* values (that is, under various pressures or suctions). Although the moisture content of a soil depends on texture, organic matter, structure and other properties, the moisture content of soils at various *pF* values follows a definite pattern. This pattern is most easily shown on a graph relating the percentage water remaining in the soil to *pF* and is called the *moisture characteristic* of the soil. An example of one of these graphs is shown in Fig. 3.3.

Two separate curves are obtained for the moisture characteristic of a soil, depending on whether the determinations are on a soil losing water (drying out) or gaining water (wetting up). The difference between these two curves is termed *hysteresis.* These two curves fail to meet because drying out is related to the neck of the pore, and wetting up to the widest part of the pore. Realising that *pF* is a logarithmic scale (the curve in Fig. 3.3 would be even steeper if an arithmetical scale were used instead of the *pF* scale), the student will notice that soils lose their superfluous water very easily and then it becomes increasingly difficult to remove more water from them. Some of the points on the moisture curve in Fig. 3.3 (for drying out) are discussed in the following sections.

Many laboratory experiments are done on *oven-dry* soil. 'Oven-dry' is arbitrarily assumed to occur (in Australia) at *pF* 7.0 or 10,000 atmospheres pressure which is defined as the amount of water remaining in a soil after it has been dried in an oven at 105°C for 8 to 10 hours. Actually, some very firmly held film water still remains in the soil at *pF* 7.0 because, if the soil is dried at even higher temperatures, it still loses weight.

The *total moisture content* of a soil is measured by weighing a sample before and after oven-drying, then expressing the loss in weight in terms of the oven-dry weight—not in terms of the wet weight, for this is very variable whereas the oven-dry weight is reproducible. Thus:

$$\text{Total moisture content} = \frac{\text{Weight of Wet Soil} - \text{Weight of dry soil}}{\text{Weight of Dry Soil}} \times 100$$

The total moisture content usually refers to *pF* 7.0.

Water between *pF* 7.0 and 4.2 is generally referred to as *hygroscopic water*, implying that most of it is adsorbed rather than interstitial. But this is probably not so as there could be some very firmly held interstitial water between the very small soil particles. The soil in Fig. 3.3 contains about 2% moisture at *pF* 7.0 and 8.5% moisture at *pF* 4.2. Between these figures, the water is unavailable to plants; however, when compared to the atmosphere, the soil is quite damp. 50% *relative humidity* corresponds to about *pF* 6.0 and 96% relative humidity to about *pF* 4.8. Thus plants require far moister conditions than do animals or humans; for us, 96% relative humidity is most uncomfortable, but soil of this humidity is too dry for plants.

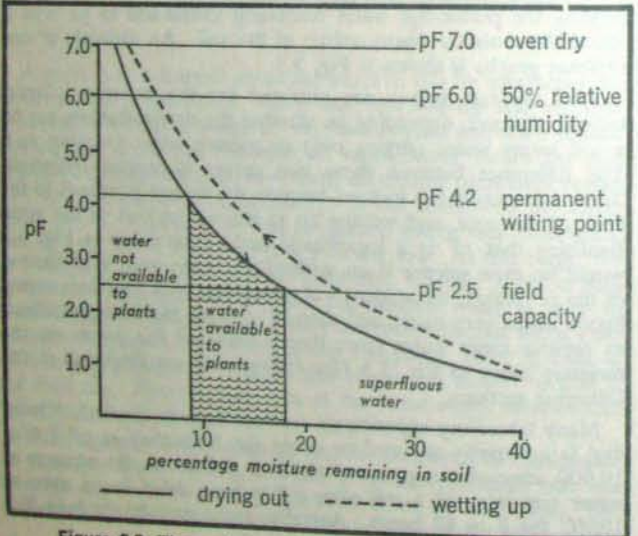

Figure 3.3: The moisture characteristic of a typical soil.

The *permanent wilting point* is generally taken as *pF* 4.2, but it varies from soil to soil and from plant to plant. It corresponds to about 15 atmospheres pressure. A plant has three or four temporary wiltings before the permanent one when it no longer revives after water is added. Thus permanent wilting point denotes the amount of water remaining in the soil when the plant wilts and will not respond to added water. Since the water in the soil at this stage is fairly immobile, it is probable that the water conditions in the immediate vicinity of the plant roots are less favourable than elsewhere in the soil. Plants may extract water even below the wilting point, provided their roots come into direct contact with the moisture film.

Permanent wilting point can be demonstrated by allowing the soil in a pot plant to dry out until the plant wilts. The plant can then be revived; then the drying and re-wetting can be repeated until the plant fails to respond. This is permanent wilting point. Next, a sample of soil can be weighed; then it can be oven-dried and re-weighed to calculate the amount of water remaining at permanent wilting point. The loss of weight of the soil will be found to be appreciable.

There is a limit to the amount of water a freely drained soil can hold. This limit is termed *field capacity* and is defined in Australia as the amount of water remaining in a soil after it has drained for 48 hours after a thorough wetting. This amount of water is a definite and reproducible percentage. When a soil is to be collected at field capacity, evaporation is prevented by covering the soil with a hessian bag or plastic sheeting for the two days after wetting; a sample is then collected in a sealed container. The sample may then be weighed, dried and re-weighed. At field capacity, many of the larger pores contain water. Field capacity varies from soil to soil, but occurs at approximately *pF* 2.5 or 1/3 of an atmosphere pressure. When a soil is drying, at field capacity, the downward movement decreases substantially and loss by upward movement (capillarity and evaporation) becomes significant. Field capacity is also termed *capillary capacity*, as the capillary pores are believed to be filled at this stage.

Any water in the soil in excess of field capacity is termed *gravitational or superfluous water* as it is easily removed by drainage. Soils do not normally remain wetter than field capacity for any length of time; indeed, to remain wetter there must be continuous rain or at least heavy falls every second day. Plants therefore experience only a small range of the soil's moisture characteristic (see Fig. 3.3), the range between field capacity and permanent wilting point.

Sticky point, which is used in the field determination of texture (see Chapter 2), approximates field capacity. The soil is mixed with water to a muddy consistency and then dry soil is added until the mixture just adheres to the fingers.

MOVEMENT OF WATER THROUGH SOILS

Water moves through soils either as a liquid or as a vapour. Movement as a liquid is caused by gravity or capillary pull, either alone or in combination. Since gravity movement occurs only in very wet soils, it is restricted to the larger pores. Water moves rapidly (at the rate of several inches a day) only through soils containing large pores, such as sands. Rapid water movement requires pores at least 1/20 mm wide. In heavier soils, once the clay is wet, swelling may block the pores and thus greatly restrict the movement of water.

Movement from a wet soil to a moist or dry soil is relatively rapid, that is, from a soil above field capacity to a soil below field capacity; but the movement of water from a moist soil to a dry soil is negligible. So if a soil is dry, water should be given directly to the root, as one cannot expect water to move in from a nearby moist zone. Plants have been observed to die from too little water when moist soil was only a few feet away.

Capillary movement may take place in all directions, but downward movement is slightly faster than upward. The height of rise of water in a soil by capillarity is inversely proportional to the diameter of the pore: as shown roughly by the following equation:

$$d = \frac{0.3}{h}$$

where d is the diameter of the pore in cm and h is the height of the rise in cm.

Thus water held at pF 3 is in pores up to 3 microns across; or, if the pores are 3 microns across, water will rise by capillarity to a height of 1,000 cm or approximately 30 feet. Water held at pF 2 is in pores up to 30 microns wide, or, if the pores are 30 microns wide, water will rise by capillarity to a height of 100 cm or approximately 3 feet.[2]

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[2] 1 micron is 0.001 mm or 0.0001 cm. The first calculation is:
$d = \frac{0.3}{1,000} = 0.0003$ cm or 0.003 mm or 3 microns.
There are roughly 30 cm to the foot, so 1,000 cm roughly equals 30 feet. The second calculation is:
$d = \frac{0.3}{100} = 0.003$ cm or 0.03 mm or 30 microns.
These calculations ignore several important variables, and so are only approximations. Water would probably not rise as far as these calculations suggest.

These simple, rough calculations demonstrate that capillary rise is much greater in heavy textured soils than in sands where the pores are much larger. The *capillary fringe* is the height above the water to which the capillary pores remain saturated, and water in this zone is at pF less than 2.0.

The upward movement of water by capillarity under conditions permitting free evaporation is not nearly as significant as was previously believed. Upward capillary flow is effective initially until the water table has been lowered substantially. It then tails off to a negligible amount. Emphasis in research is now concentrated on the limitations of capillary flow in explaining the losses of water at the surface of soils by evaporation.

Capillary movement is always from regions where water is loosely held (low pF values) to regions where water is tightly held (high pF values), so that the tendency is to equalise energy.

Any movement of water within the soil when capillarity can no longer operate (when the soil is too dry) must be by vapour pressure differences within the soil. Since, even at wilting point, the soil atmosphere is practically saturated with water, movements of water in the vapour phase are important only at pF values higher than 4.2. When the soil is drier than wilting point, water vapour movements are enhanced by temperature fluctuations, for example downwards in the afternoon from warmer to cooler horizons and upwards in the morning when the topsoil is cooler. This movement accounts for a considerable loss of moisture from the soil, especially in warm weather, causing the topsoil to dry below wilting point for several inches.

In addition to losses of water from the soil by drainage, and by capillary and vapour movement followed by evaporation, a great deal of water is lost by *transpiration*. This is the mechanism whereby plant roots absorb water which moves through the plant distributing the essential nutrients and is eventually lost to the atmosphere through evaporation, mainly through the stomata, the apertures for respiration in the leaves. Transpiration losses may be enormous. The loss is generally expressed as the *transpiration ratio or coefficient*, which is the units of water transpired by the plant in the production of every unit of above-ground dry matter. The transpiration ratio varies with the amount of water available, the period of the day when the plant's stomata are open, and the solar energy falling on the crop. A transpiration ratio of 500, to produce 4,000 lbs of dry matter per acre, involves a water loss from the soil equivalent to nine inches of rainfall during the growing season. Lucerne, clover and most grasses have transpiration ratios near 800, potatoes, oats and cow-peas near 600, wheat and barley near 500, maize and sugar beet near 400 and millet and sorghum near 300.

It is true that a vegetative cover greatly reduces the rate of free evaporation of water from the soil surface, but transpiration losses more than compensate for this. *Fallowing* is a well-known practice to conserve soil water by reducing transpiration losses. In the Australian wheat belt, where fallowing is widely practised, it has been claimed that the moisture conserved may be the equivalent of three inches of rainfall, which is appreciable when the fall in the growing season may be only 10 inches. It was previously believed that the cultivation of fallow was effective in conserving water because it broke the soil capillary channels and thus checked losses by free evaporation; however, it is now thought that the main effect lies in the removal of weeds and the checking of transpiration losses. Removal of weeds without harrowing has proved almost as effective as harrowing; but the cultivation of heavy clayey soils that are lying fallow does avert a loss of moisture from the lower layers because it prevents subsoil cracking from taking place.

Mulching aims to check free evaporation from the bare soil surface. A mulch is some type of thin covering over the soil surface. A mulch is not ploughed in but remains on the top of the soil to seal off the capillaries and prevent water losses. Many artificial mulches are in use. One of the most effective is straw chopped to about 5 cm in size. A thin covering of straw may reduce water losses from the soil to 1/10 the usual amount. Straw that is chopped more finely than this is less effective as it is soon incorporated into the soil. In Australia, bitumen mulches have been used with some success, although their black colour tends to raise the temperature of the soil.

Self-mulching soils are prized agriculturally. They have, on drying, a natural ability to develop a crumb structure in the upper half- to quarter-inch. This natural mulch prevents subsoil cracking and breaks the capillaries as shown in Fig. 3.4. Many Australian basaltic soils, especially the black earths and krasnozems, are self-mulching.

There are many soil problems associated with irrigation. Some soils in arid regions have first to be reclaimed before they are suitable for irrigation. Saline soils (soils containing too much sodium) may be effectively reclaimed by irrigation and the addition of gypsum (calcium sulphate), from which the calcium replaces the sodium in the soil, and re-irrigation leaches the resulting sodium sulphate. Lime is less effective in replacing sodium in soils (see Chapter 11). Repeated irrigation may eventually raise the *water table* (the level below which the soil is

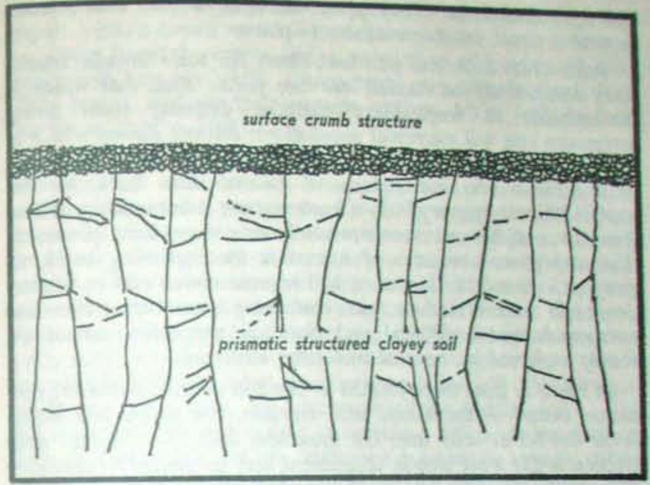

Figure 3.4: Cross-section through a self-mulching soil.

permanently saturated), especially in very heavy soils where drainage is impeded. Many crops are intolerant of a high water table. A water table raised to within three feet from the surface for crops (except rice) and two feet from the surface for pasture, is usually dangerous. The raising of the water table by irrigation, or by heavy rainfall in an exceptionally wet season, may bring salting problems as some ground water is more saline than the water above. High water tables must be lowered by drains, and the deeper the drains the more quickly the water will flow into them.

SOIL ORGANIC MATTER

Soil organic matter includes all non-mineral carbon compounds found in the soil. Lime (calcium carbonate) is excluded as its origin need not be organic. To the pedologist, organic matter is vital because soil horizons probably do not develop until organic matter is present. Organic matter is the main pigment responsible for black and brown colours in soil. It has profoundly beneficial effects on the soil's physical properties: it encourages granulation; it makes very heavy soils easier to work by reducing plasticity and cohesion; and it increases the water-holding capacity of the soil (organic colloids have approximately three times the water-holding capacity of inorganic colloids). In addition, organic matter acts as a storehouse for many plant nutrients, which, after certain reactions occur, become available to plants.

Aside from fresh leaf and root litter, the soil's organic matter may conveniently be divided into two parts: first, that which is recognisable as fragmentary material deriving from living organisms (the soil *microbial population*); second, *humus* which is unorganised, darker in colour and gel-like in properties. Although investigations into the nature of humus date back to the early 19th century, very little is known about it because crystalline derivatives of the main constituents have not yet been prepared. The amorphous character of humus is the stumbling block to progress in this field. However, soil organic matter can be broken down into several fractions, each containing a number of chemical compounds whose physical and chemical properties cannot be simply explained in terms of molecular structure.

In Fig. 3.5, from three-fourths to one-half of the soluble organic matter occurs in the humic acid fraction, the proportion being lower for forest soils than for grassland soils. The fulvic acid fraction is the most mobile constituent and so generally increases with depth down the soil profile. Fulvic acid appears to be the potent factor in the process of podzolisation, and may well be the major complexing agent assisting the translocation of iron and aluminium down the profile. Fulvic acid, unlike humic acid, is colourless or only slightly coloured, so its accumulation does not impart a black colour to the soil. Soils which are black in colour have been found to contain higher proportions of humic acid than soils which are not black. Chernozems may contain more than 35% humic acid in their organic matter, while podzols may contain as little as 10%.

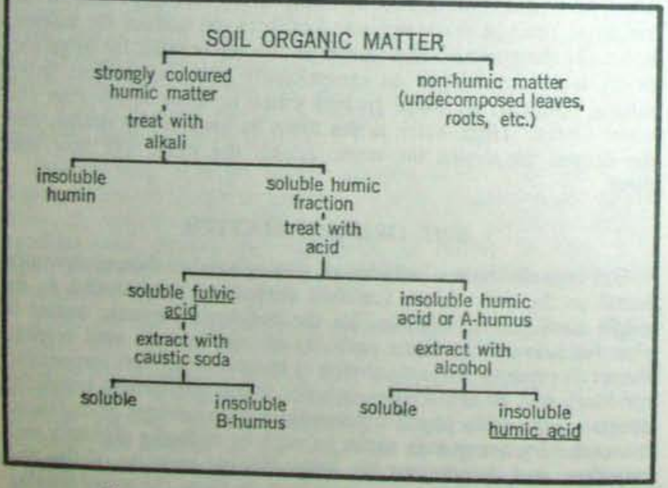

Figure 3.5: Fractionation of soil organic matter.

An important difference between soils and newly weathered rock is the fact that soils contain a living population of micro-organisms which obtain their energy by oxidising organic residues left behind by plants and the animals that eat these plants. In turn, the plants growing in the soil subsist on the products of microbial activity. Micro-organisms in the soil may be divided into *microflora* and *microfauna*. *Microflora* include bacteria which are single cell plants (perhaps numbering 4,000 million per gram of soil), fungi which require moist conditions, actinomycetes which can survive in dry soils but cease to grow below *pH* 5.0, and algae which are single cells, the precursors of the higher plants. The soil *microfauna* consist mainly of protozoa, which are the simplest form of animal life and are important in reducing organic debris to its elemental constituents, and in making nitrogen available to plants. The invertebrates (worms, termites and arthropods) are important in moving humic material through the soil and in reducing it to smaller dimensions. Earthworms, which are found in moist soils, cut up and ingest large amounts of plant material which is humified in their intestinal tracts. Their excreta, which are neutralised by lime secretions, can be consumed by smaller organisms. Worms, ants and termites also favourably influence soil structure and promote aeration.

Total soil organic matter may be estimated as organic carbon (organic matter contains approximately 57% carbon). Organic carbon may be estimated by digesting the soil in acid potassium dichromate and the quantity of dichromate used calculated by titration against ferrous sulphate. The equipment needed for estimating organic carbon is fairly elaborate, so estimations are usually carried out only in soils laboratories.

Soils that are extremely high in organic matter are called *peats*. The amount of organic matter required for classifying a soil as a peat varies among pedologists, but it is always very high, usually above 40%; even if a soil is very black but contains less than this amount it will not be a peat. The high amount of organic matter in peaty soils produces a friable consistency, making sandy soils more coherent and clayey soils less plastic. Peaty soils are often associated with a slowly rising water table, or steady sedimentation of a lagoon, both conditions having rush and sedge vegetation. In Australia, peaty soils are found in infilled lagoons of coastal barrier systems and in ill-drained areas in highland regions such as the Snowy Mountains. Poorly drained glaciated regions of Northern Eurasia and North America have well-developed peaty soils. The high concentration of organic matter prevents podzolisation, though rainfall may be ample.

SOIL MINERAL MATTER

Soil mineral matter is that part of the soil which is most clearly affiliated with the soil's parent material. Soil air and soil water are closely related to the soil's external environment, while soil organic matter is related to the vegetation growing in the soil, but most of the soil's mineral component is derived from the underlying parent material. Some mineral matter may be added externally in soils receiving steady accessions from above, for example soils on the lee side of deserts receiving additions of loess, or alluvial soils receiving periodic topdressings of silt after floods. In most soils, however, the mineral component is wholly derived from the underlying parent material, the material which has been transformed into the soil horizons.

Soil mineral matter may be divided into *primary* and *secondary* components. To a certain extent this division is arbitrary, as minerals which are identical in appearance to primary rock minerals may form authigenically in the soil profile (an example is apatite which can form from soil phosphates and may be identical to primary, igneous apatite); conversely, not all so-called secondary minerals are the products of weathering in the soil itself. Thus there are many ways of defining the terms primary and secondary. One way is to regard primary minerals as those which have not undergone chemical change during the period of soil formation that is being considered. Primary, in this sense, implies residual; however, it is extremely difficult to distinguish between the primary and secondary minerals in soils formed from parent materials containing minerals which have weathered subaerially in a previous cycle: that is, most of the sedimentary and some of the igneous and metamorphic rocks.

A more practical distinction between primary and secondary minerals reserves the term primary for those minerals which have never weathered subaerially and so excludes many of the minerals of sedimentary rocks. But not all igneous rocks contain minerals which are chemically identical to the form in which they first crystallised from molten magma. When magma consolidates, certain minerals crystallise out first, leaving the remaining magma deficient in the components from which they formed. Late magmatic solutions may be chemically incompatible with these early minerals and may attack their surfaces, altering them to clay

minerals. Such alterations, which occur during the later stages of consolidation of the magma and therefore at quite high temperatures, may resemble the subaerial alteration products of the same mineral. An example may be taken from the consolidation of doleritic magma. Olivine is one of the first minerals to crystallise at the initially high temperatures. It is attacked by late magmatic solutions to form a group of clay minerals which may be reddish in colour (from the iron in the olivine). This alteration product is termed *iddingsite* and contains the same group of clay minerals which can form from the subaerial alteration of olivine. Alteration of primary minerals by late magmatic solutions at high temperatures is referred to as *deuteric alteration*. Thus, although igneous rocks may have never weathered subaerially they may contain clay minerals like those which form from subaerial alteration. Because of the difficulty of distinguishing minerals altered deuterically from those altered subaerially, some pedologists prefer to reserve the term primary for minerals unaltered both deuterically and subaerially. This is the sense in which the term primary is used in this book.

Primary minerals are unaltered chemically from their initial form when the parent magma first consolidated. They comprise the greater part of igneous and metamorphic rocks and some sedimentary rocks whose primary minerals may have survived several cycles of attack; quartz, for example, is an extremely resistant primary mineral and is the main constituent of sandstones. *Secondary* minerals are the alteration products of primary minerals. They may have altered *in situ* through the loss of certain constituents; but, more often, they represent the combination of minute fragments weathered from primary minerals. These fragments may be ions (charged elements, e.g., Na^+) or small structural units which carry a charge. Because secondary minerals are weathering products, they are usually smaller in size than primary minerals.

Because primary minerals tend to be larger than secondary minerals, there is a correlation between the nature of minerals and the size fractions of the soil. The sand fraction (2.0—0.02 mm diameter) tends to contain primary minerals, the clay fraction (less than 0.002 mm diameter) tends to contain secondary minerals, and the silt fraction (0.02—0.0002 mm diameter) contains both primary and secondary minerals. This concentration into fractions is incomplete because colloidal (clay-sized) primary minerals are known, especially quartz, and the sand fractions may contain aggregates of secondary minerals which develop either through the alteration *in situ* of primary minerals (e.g., iddingsite) or through the very firm cementation of groups of secondary minerals which cannot be separated by mechanical analysis (e.g., the iron oxides, many of which are counted in the sand fractions).

The minerals found in soil may be divided into the following groups:

1. *Primary aluminosilicates.* These make up the greater part of the sand fraction of most soils and are the parent materials for the secondary minerals.

2. *Silicate clays.* These are secondary minerals and make up the greater part of the clay fraction of many soils.

3. *Sesquioxides.* These are secondary minerals and consist of the hydrated oxides of iron, aluminium, manganese and titanium. They are found in both the sand and clay fractions and are the main constituent of tropical soils.

4. *Other oxides and salts.* These include the soluble and slightly soluble materials which are neither silicates nor sesquioxides. They may be either primary or secondary minerals. They are important in the soils of desert regions and include lime, gypsum and halite (sodium chloride).

1. Primary Aluminosilicates

Most of the soil's primary minerals are aluminosilicates; far less important are oxide, sulphide and carbonate primary minerals. Silicon is just as important to the inorganic world as carbon is to the organic world. Carbon and silicon have certain affinities; they belong to the same group in the chemical periodic table and have four-fold co-ordination. Silicon has the remarkable ability to form linkages with oxygen atoms and other silicon atoms, while aluminium and other cations occupy the interstices formed by these linkages. The aluminosilicate minerals are complex both physically and chemically. Their physical structure may be exceedingly complicated three dimensional frameworks of silicon and oxygen sheets linked by various cations which may replace each other. Chemically, they are even more complicated: certain positions in the structure are weaker than others so that cations and anions may enter or leave the structure in either case, causing the breakdown which constitutes weathering. The relative ease of weathering of the aluminosilicates depends on many variables, both internal and external to the mineral. The most important internal variables are (a) the nature of the cations holding the structures together, (b) whether the structures are electrostatically neutral, and (c) whether there are empty positions in the structure through which water can penetrate.

The aluminosilicate minerals may be classified according to their chemical composition and structure as follows:

1. *Orthosilicates,* which are independent groups of silicon and oxygen linked by basic cations. In the olivine group, the linking cations are iron and magnesium. Iron is readily oxidised in the presence of water, causing olivine to weather very readily to a reddish mineral, although the structure of olivine is a fairly stable one. Other members of this group, such as zircon which is linked by the relatively inert element zirconium, are much more resistant to weathering.

2. *Chain structures with single or double chain linkages.* The pyroxenes have single chains of silicon, aluminium and oxygen, linked by magnesium, iron and calcium and are fairly easily weathered, losing their cations. The amphiboles have double chains of silicon, aluminium and oxygen, linked by various cations. The stronger the linkages of the silicon, aluminium and oxygen, the more resistant the mineral is to weathering, so the double chain minerals tend to be more resistant to weathering than single chain minerals.

3. *Sheet structures.* The amphibole (e.g., hornblende) chain can be envisaged as arising by the parallel linking of pyroxene (e.g., augite) chains. If this process were continued indefinitely, a sheet structure would result. The micas, chlorites, and silicate clays all have this basic sheet structure. The sheets are held together by various cations including potassium, which provides a weak point in the structure and is relatively easily lost. There are two main types of mica: biotite or black mica, and muscovite or white mica, which is more resistant to weathering than is black mica.

4. *Three dimensional structures.* These form continuous frameworks of linked silicon, aluminium and oxygen atoms. The feldspars are an important member of this group. There are two main types of feldspar, orthoclase or potassium feldspar and the plagioclases, or sodium-calcium feldspars which form a series from albite (the sodium rich member) to anorthite (the calcium rich member); intermediate members contain both sodium and calcium (e.g., andesine). Feldspathoids form in rocks where there is a deficiency in silica. The feldspathoids (e.g., nepheline) weather very rapidly in soils.

Pure silicon oxide (silica) exists in many forms, the most common of which is quartz. This mineral, being a closely locked structure and lacking the cations (iron, aluminium, etc.) which make all the other minerals so vulnerable, is very resistant to weathering, although it may be chemically attacked under special conditions which will be discussed in Chapter 7.

In igneous rocks (rocks which have crystallised from molten magma) there is a definite order in which the minerals form as the magma cools. This order, which is called Bowen's Reaction Series, closely follows the list above, in that the orthosilicates crystallise out first, then the single chain silicates, then the double chain silicates, then the sheet minerals, and finally the three dimensional minerals. Thus olivine is one of the first minerals to crystallise and feldspars are one of the last. It was previously believed that Bowen's Reaction Series closely correlated with the order of subaerial weathering of minerals, and that the resistance of minerals to weathering closely followed the order of the series, but it seems that factors other than chemical composition and structure are equally important in mineral weathering. Macro-structure of minerals (whether they are prismatic, platy, rod-shaped, etc.) seems to be very important, and some of the platy minerals (such as biotite) appear to weather more rapidly than they should according to their position in the reaction series. Bowen's Reaction Series, for the minerals discussed so far, may be written as shown in Fig. 3.6.

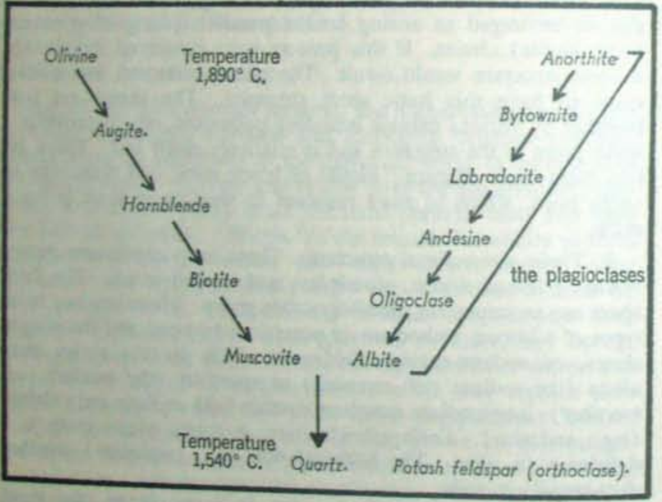

Figure 3.6: Bowen's reaction series.

The rate of weathering of some minerals follows this order; for instance, olivine is one of the most readily weathered minerals and quartz is one of the most stable under various soil conditions. The calcium plagioclases (anorthite, bytownite, etc.) are more easily weathered than the sodium plagioclases (albite, oligoclase, etc.), while orthoclase (potash feldspar) is the most resistant of all the feldspars. Although orthoclase crystallises at about the same temperature as quartz, at the end of the consolidation of the magma, quartz is far more resistant in soils.

Minerals such as quartz, which resist weathering, tend to remain in the sand fraction of soils, while the readily weathered minerals, such as olivine and pyroxene, are soon lost and their weathering products contribute to the soil's clay fraction. The readily weathered minerals dominate only in the *C* horizons of soils derived from the underlying rock. They may be found in the *B* horizons, but are less common in the *A* and *E* horizons where the resistant minerals tend to accumulate.

Thus the particle size distribution in a soil is closely related to the parent material. Where this is rich in resistant minerals, the soil tends to be sandy; where it lacks resistant minerals, the soil tends to be clayey. Since rocks are classified according to their constituent minerals, and mineral weathering strongly influences most soil properties, it is important for pedologists to understand the mode of occurrence of minerals in rocks.

Igneous rocks may be grouped into *acid* rocks which are rich in silica (this has nothing to do with acid in the sense of high hydrogen ion concentration) and *basic* rocks which are poor in silica. Basic rocks tend to contain the high temperature minerals (indicated in Bowen's Reaction Series) as these are poorest in silica, while acid rocks contain the low temperature minerals as these are richest in silica. The high temperature minerals are found together in basic rocks, while the low temperature minerals are found together in acid rocks. These relationships are shown in Fig. 3.7. It is rare to find a rock containing, for example, olivine and quartz, olivine and orthoclase, or olivine and biotite. Where mica is found in basic rocks, it is quite different chemically from ordinary biotite.

From Fig. 3.7 one can learn a great deal about soil formation. The rocks on the right hand side of the table tend to form clayey soils (they contain readily weathered minerals), while the rocks on the left hand side tend to form sandy soils. Since the minerals on the right hand side are higher in cations (iron, magnesium, calcium, etc.), soils derived from rocks containing them tend to be more productive. Thus a basaltic soil is both more productive and more heavily textured than a granitic soil.

The nature of the aluminosilicate minerals found in soils is related, therefore, not only to their weatherability, but to the parent material. Parent material as a soil-forming factor is dealt with in Chapter 5, where sedimentary and metamorphic rocks are also discussed. Fig. 3.7 demonstrates the relationship between the

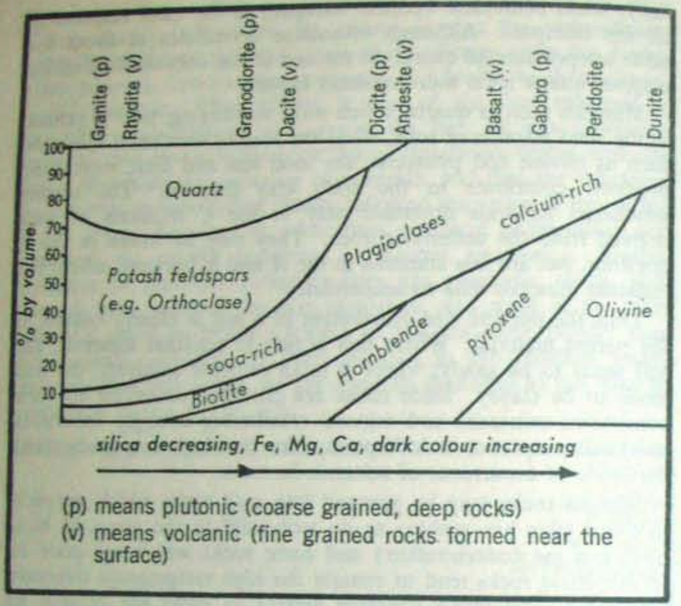

Figure 3.7: Mineral composition of common igneous rocks.

common minerals and rock types. From Fig. 3.7, it is obvious that granitic rocks contain quartz, much orthoclase, a little plagioclase, hornblende, and sometimes biotite. Basaltic rocks contain olivine, plagioclase and pyroxene.

The Weathering of Aluminosilicate Minerals

Weathering is conveniently divided into physical and chemical. *Physical weathering* involves the breaking of minerals into smaller fragments without chemical change. In weathering physically, minerals increase their surface area and so make themselves more susceptible to chemical attack. It would not be wise to try to assess the relative importance of physical and chemical weathering or even to attempt to separate them as the two are so interconnected that it is doubtful whether one operates without the other. Chemical weathering, which involves the loss of constituents and the formation of new minerals, is present wherever there is water. Physical weathering operates wherever there is sufficient pressure upon minerals to cause them to shatter. This pressure may arise from internal conditions within the mineral, such as

chemical zoning or the presence of voids or inclusions, or it may result from external conditions, such as temperature fluctuations, root growth, animal burrowing, frost shattering, or constriction when the enclosing soil swells or shrinks. The importance of physical weathering lies not in the actual reduction in the size of the mineral concerned, but in its increased potential for chemical attack as a result of its increased surface area. The greater the surface area of a mineral, the higher are the number of possible points for chemical attack. At the surface of a mineral, weak bonds, vulnerable cations and empty positions in the structure are exposed.

Even in desert areas, chemical weathering is vital, though the rate may be slower than in humid areas because of lack of water. Water is essential to chemical weathering, because the weathering reactions are ionic and ions exist only in solution. *Chemical weathering* is essentially the movement and replacement of ions. When ions leave a mineral or are replaced by smaller ions, the structure may collapse, releasing more ions. The categorising of weathering reactions as 'hydration', 'hydrolosis', 'hydroxylation' as is done in many pedology textbooks involves cumbersome terminology and serves little purpose, so it will not be attempted here. More important is an outline of some of the ionic reactions which occur during chemical weathering.

The aluminosilicate minerals may be regarded as sheets of oxygen atoms held together by cations (silicon, aluminium, calcium, potassium, magnesium, iron, etc.). Oxygen anions have a radius of 1.32 Å[1] and are roughly spherical in shape. Spheres, when packed so as to occupy the smallest possible volume, contain two types of interstice or space between them. This may be demonstrated by taking tennis balls, representing the oxygen atoms, and packing them in rows so that they occupy the smallest volume. It will be noticed that there is a basic reproducible pattern and that the interstices between the balls are of two sizes. One type of interstice is surrounded by four balls (this is called the tetrahedral position), and the other is surrounded by eight balls (this is called the octahedral position); the latter interstice is clearly larger than the former. The two positions are illustrated in Fig. 3.8, which shows the positions in plan and section, with the balls of another plane dotted in. The balls surrounding the tetrahedral position form a tetrahedron, which, for oxygen atoms, has a diameter of 0.30 Å. The octahedral position is surrounded by six balls, with one at the top and another at the bottom a long way from the octahedral interstice. These top and bottom balls are

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[1] The Angstrom Unit (Å), which is used to measure the radius of atoms, is 10^{-8} centimetres or 0.00000001 cm.

shown by light lines in Fig. 3.8 (d). The diameter of the octahedral interstice is 0.70 Å for oxygen atoms.

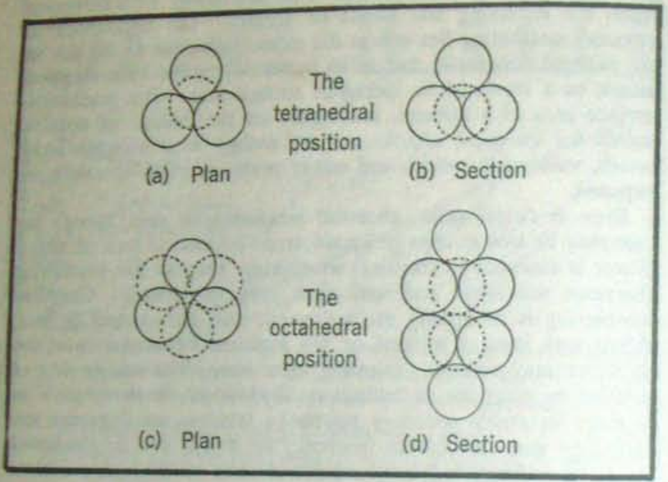

Figure 3.8: Tetrahedral and octahedral positions in aluminosilicate minerals.

Because the tetrahedral and octahedral interstices are quite small relative to the size of cations which could hold the oxygen atoms together, only a few cations fit the interstices. The tetrahedral interstice is so small that only silicon (diameter 0.39 Å) can fit it without undue strain, as the oxygen atoms are not rigid and so will bend a little. Aluminium (diameter 0.57 Å) which is also found in the tetrahedral position is much too large and causes strong distortion of the structure, making it more prone to collapse by weathering. Aluminium, on the other hand, is only a little too small for the octahedral position, where it fits with little strain. The other cations found in the octahedral position are too large and thus are vulnerable to chemical attack. The common cations found in the octahedral position are:

lithium	0.78 Å diameter	magnesium	0.78	ferrous (iron II)	0.83
sodium	0.98	calcium	1.06		
potassium	1.33	barium	1.43	ferric (iron III)	0.67
ammonium	1.43	manganese	0.91		

Thus one way in which minerals weather is by breakdown at weak points caused by the presence of cations that are either too small or too large to fit the interstices they occupy. Bonds are broken and the cations released.

Not every interstice is occupied by a cation, so cations from the soil solution may diffuse into the structure to occupy the empty positions. This may upset the electrostatic neutrality of the structure and force other cations to leave, causing the structure to collapse. Since water contains hydrogen ions in the concentration 10^{-7} moles per litre, even if the soil is neutral or alkaline, hydrogen ions are available to take part in ionic reactions. In acid soils, there is an even higher concentration of hydrogen ions; and in alkaline soils, there is a higher concentration of hydroxyl ions. Since the hydrogen ion is the smallest of the cations, it can occupy even the smallest position in the structures. The water molecule, being a dipole, behaves not unlike a cation and can fit into structures, upsetting their electrostatic neutrality. The hydronium ion (H_3O^+) is not much larger than the water molecule, is fairly stable in soil solutions, and acts as a retarding influence on the active hydrogen ion. The entry of water or ions from water (hydrogen, hydronium or hydroxyl) into minerals is a part of the process of chemical weathering.

Another important weathering reaction is the oxidation of ferrous (iron II) to ferric (iron III) ion by the removal of one electron. The latter ion is smaller than the former and, because it carries one more positive charge, it promotes structural instability. Thus minerals containing ferrous (iron II) ion have a weakness.

A single weathering reaction may be considered from several viewpoints. Consider the following weathering reaction:

$$\text{K biotite} + H_2O \longrightarrow \text{H biotite} + KOH$$

or, ionically

$$\text{K biotite} + H^+ \longrightarrow \text{H biotite} + K^+$$

This reaction, which involves the loss of potassium from the biotite, may be regarded as *oxidation* (K forms K^+ by the loss of an electron), *hydration* (H^+ enters the structure), a *cation exchange reaction* (K^+ is replaced by H^+) or a *diffusion* reaction (H^+ diffuses into the empty position left by the departure of K^+). In fact, this reaction may be labelled by all of these terms. It is more important to consider the reactions in terms of the ions involved than to describe them by a chemical term.

2. THE SILICATE CLAYS

From the above considerations, it is apparent that weathering reactions produce cations and anions and broken fragments of silicate minerals (groups of cations enclosed by four or six oxygen atoms). One of these small fragments may act as a nucleus and attract to it other charged bodies suspended or dissolved in the soil solution. In this way, a clay mineral is born from the products of weathering of the aluminosilicate minerals. Since aluminium in the tetrahedral position is unstable, the aluminium in clay minerals is more likely to be in octahedral co-ordination. Silicon, on the other hand, is more likely to be in the tetrahedral position; so clay minerals are likely to contain silicon tetrahedra and aluminium octahedra. The nature of the clay mineral so formed depends on the reactants present as well as on the environment of formation (in particular, the *pH*).

The silicate clays build up, little by little, about the nucleus as more reactants become available and conditions favour their precipitation. Most silicate clays are crystalline (a few more recently examined types such as allophane are non-crystalline) with structures resembling those of the sheet aluminosilicate minerals (the micas) already discussed. The silicate clays are plate-like in structure under the electron microscope. Because of the excess oxygen in their sheets, their charge is negative, so they behave like huge *anions*, attracting cations and the positive ends of water molecules as shown in the diagram (Fig. 3.9).

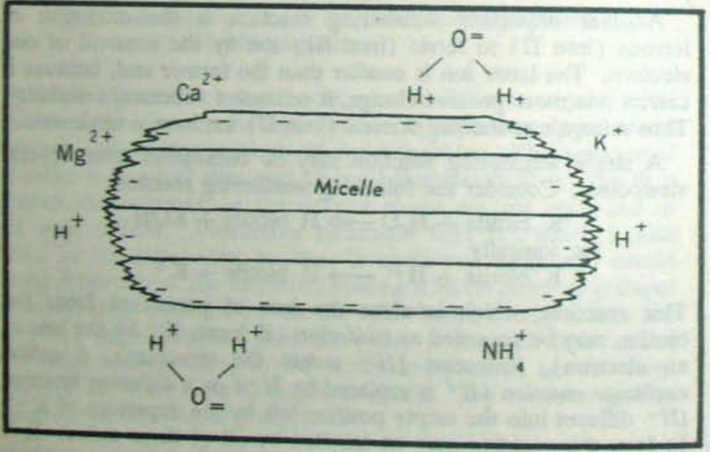

Figure 3.9: Diagrammatic representation of a clay mineral.

The negative charge on the nucleus, which is called the *micelle*, may be demonstrated by submitting suspended clays to an electric current—the micelle moves towards the positive electrode. Since the cations adsorbed onto the micelle are only electrostatically held, they are readily removed or replaced by other cations. This replacement of cations on the micelle by other cations is termed *cation exchange*, and the ability of the clay to exchange cations is termed the *cation exchange capacity* (*C.E.C.*). As the cations are held rather loosely, they may all be exchanged. In the laboratory, one can replace all the cations by ammonium (NH_4^+) by leaching the soil with ammonium chloride. In the field, when a farmer adds gypsum ($CaSO_4 \cdot 2H_2O$) to reclaim a sodium-saturated soil, the calcium replaces the sodium as shown:

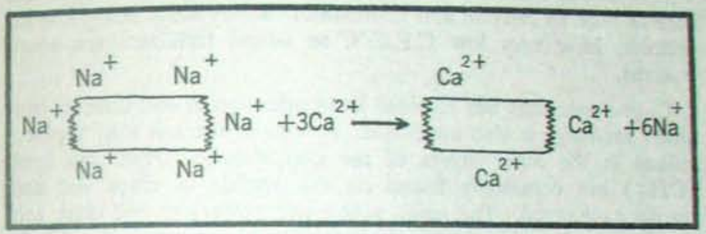

Figure 3.10: Diagrammatic representation of cation exchange.

The student will notice that since calcium carries two positive charges, it replaces two sodium ions. In nature, the efficiency with which ions replace each other depends on the ion concentration in the soil solution, the activity of the ions, and the number of charges on them. Ions which attract water are greatly reduced in their speed of travel (lithium is a commonly hydrated ion). Taking these and other variables into account, the common soil ions may be rated as follows in descending order of their replacing power:

$$H > Ba > Ca > Mg > K > NH_4 > Na > Li$$

This means, for example, that sodium and lithium are very easily replaced by hydrogen. When there are similar concentrations of both sodium and hydrogen ions, hydrogen, being more active, tends to replace sodium. Thus sodium-saturated colloids are found only in desert regions or areas of low relief where the concentration of sodium is high relative to hydrogen. Soils of humid areas are likely to contain clays saturated with hydrogen. Soils of semi-arid areas may contain hydrogen and calcium-saturated clays, while desert soils may contain appreciable amounts of exchangeable sodium and potassium.

The total cation exchange capacity of soils is expressed in milli-equivalents per 100 gms of soil. This measure is related to the hydrogen ion which is monovalent. The cation exchange capacity of a soil may be determined by replacing all of the cations with ammonium and estimating the amount required for this replacement. It depends on the organic colloids[2] and sesquioxides as well as the silicate clay minerals. Organic colloids have a cation exchange capacity (C.E.C.) of 200 - 450 m.e./100 gms. The C.E.C. of the sesquioxides and silicate clays is lower, ranging up to 100 m.e./100 gms, so the C.E.C. of many soils of temperate areas is somewhere in the vicinity of 50 m.e./100 gms. A high C.E.C. means that a soil has a strong ability to fix added fertiliser cations such as calcium and ammonium. Sandy soils, being low in colloids, have very low C.E.C.'s, so added fertilisers are soon leached.

Cation exchange has received most attention in soil science, but *anion exchange* is also important. Anions in the soil may replace anions in the outer layers of the clay micelle. Hydroxyl ions (OH^-) are commonly found on the outside of clays and are readily exchanged. The anion adsorptive powers of soil clays are found to be associated with the presence of iron and aluminium compounds since the removal of these substances greatly reduces anion adsorption. Anion adsorption is particularly important in the fixation of soil nutrients, especially phosphates.

Clay minerals have the ability to adsorb water molecules, which are dipoles, so that their positive ends can attach themselves to the negative clay micelle. A clay can adsorb water from both the soil and a moist atmosphere. When a dry soil adsorbs water, heat is evolved, called the 'heat of wetting'. At low vapour pressure, clays saturated with potassium, sodium and ammonium adsorb less water than clays saturated with hydrogen and calcium; but at high vapour pressures, sodium-saturated soils adsorb the largest amount of water.

The silicate clay minerals may be classified according to the position of the silicon in relation to the oxygen sheets, into two groups, one of which has a more open structure than the other. The second, more compact group can be subdivided according to the number of silica and alumina sheets in the lattice and the degree of hydration. A classification of silicate clays is outlined in Fig. 3.11.

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[2] The term *colloid* is defined by size. Colloids are generally considered to be particles between 2 and 200 microns in size, so they are larger than the large molecules such as sucrose, yet too small to be seen under the microscope. They may be organic or inorganic. The term *clay*, also defined by size, refers to inorganic colloids.

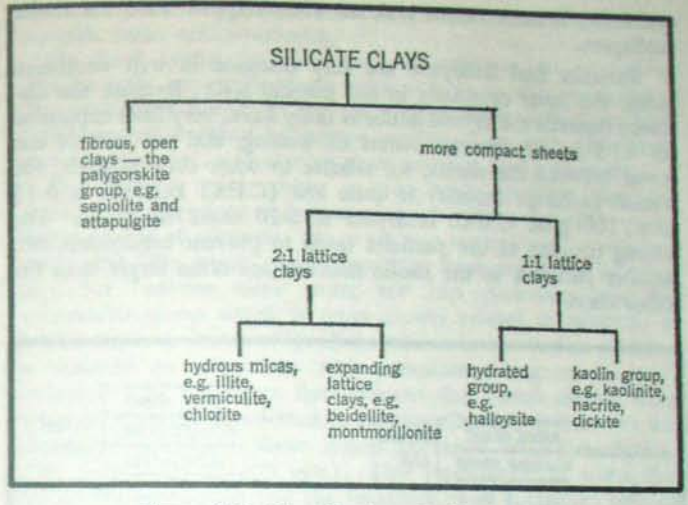

Figure 3.11: Classification of silicate clays.

The clays of the *palygorskite* group are fibrous, open and well hydrated. The minerals of this group are not common in soils, but are important in lake deposits which have salt-lagoonal characteristics. Because of this association, these minerals may indicate the origin of some transported soils. In the Riverina in New South Wales, the mineral palygorskite has been associated with one of the wind-transported soil parent materials. Beattie and Haldane (1958) consider that this mineral suggests aeolian transport from the west as the origin of the Widgelli parna, with which it is associated. Palygorskite is restricted to the Widgelli parna and is not found in any of the other allegedly wind-transported soil parent materials.

The *1 : 1 lattice clays* include the *kaolin group* (sometimes referred to as the kandites, from the minerals *kaolinite* and *dickite*) and the *hydrated 1 : 1 lattice clays.* The crystal unit of the 1 : 1 lattice clays consists of a silica sheet on top of an alumina sheet firmly bonded in octahedral co-ordination. The sheets are firmly held by –O–OH– bonds. The variations between kaolinite, nacrite and dickite relate to the stacking of the silica and alumina sheets. Halloysite has a structure similar to kaolinite, but with a layer of water hydrogen-bonded between the silica and alumina sheets. The water is firmly held in each sheet and when halloysite is dehydrated, the resulting mineral contains more spacing than kaolinite, because of the residual water trapped when the lattice collapses.

Kaolinite and halloysite are very common in well weathered soils, the latter especially in red tropical soils. Because the distance between the crystal lattice is fairly fixed, very little expansion of 1 : 1 lattice minerals occurs on wetting, and few cations can enter between the sheets; so, relative to other clay minerals, the cation exchange capacity is quite low (C.E.C. kaolinite is 3-15 m.e./100 gms, C.E.C. halloysite is 6-10 m.e./100 gms). The strong bonding of the particles tends to prevent breakdown into smaller particles, so the kaolin minerals are often larger than the other clays.

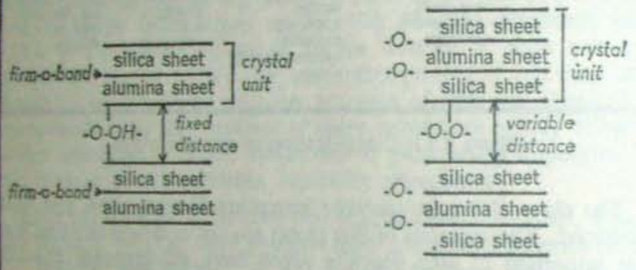

Diagram of 1:1 lattice clay, e.g. kaolinite

Diagram of 2:1 lattice clay, e.g. montmorillonite

Figure 3.12

The *2 : 1 lattice clays* include the *hydrous mica group* and the *expanding lattice group* (also referred to as the montmorillonite or smectite group). The crystal unit consists of an alumina sheet firmly bonded between two silica sheets, but, unlike the kaolin group, these units are only weakly held by –O–O– bonds. Water may freely enter the space between the crystal units, so the distance between these units does not remain fixed. Hence these clays readily expand and have high cation exchange capacities (they have internal as well as external surfaces). The C.E.C. of montmorillonite is 60-100 m.e./100 gms. Some of the aluminium in montmorillonite is replaced by iron and magnesium. This replacement by iron and magnesium causes a weakness in the structures, so that montmorillonite may alter to more stable clay minerals under acid weathering.

The 2 : 1 lattice clays are most abundant in the neutral to alkaline soils, especially under grassland, as in the chernozems. The expanding lattice of the clays gives these soils their high shrinking and swelling properties, making them very difficult to manage when wet, and causing them to crack badly when dry.

The hydrous mica group is even more unstable than the montmorillonite group and alters quite readily to one of the more stable groups of minerals. There are two main groups of hydrous micas, the *illite* group which resembles muscovite mica in structure but contains more water and less potassium, and the *vermiculite* group which is more closely related in structure to biotite mica. In illite, 15% of the silicon in tetrahedral positions is replaced by aluminium, with potassium taking up the extra valency. The potassium lies between the crystal sheets, fixing them rather more firmly than in montmorillonite, so illites are less expansive and have lower cation exchange capacities (C.E.C. illite is 10-40 m.e./100 gms). This replacement of tetrahedral silicon for aluminium and the introduction of potassium between the lattice weakens the structure, so illites weather quite readily. Vermiculite crystal units are not so firmly bonded, so the cation exchange capacity is higher than in the illite group (C.E.C. vermiculite 150 m.e./100 gms).

When micas weather in soils, potassium and the other cations slowly diffuse out of the structures. Potassium release is hastened by the removal of potassium by plants and by leaching, and water moves into the lattice to form the hydrous micas. Under slow potassium release (weak leaching) the weaker hydrous micas form; and under rapid potassium release (more intense leaching) the more stable hydrous micas form; and finally montmorillonite forms. Thus the hydrous micas may be transitional minerals formed under initially weak leaching, as shown in this equation:

$$\text{Mica} \rightleftharpoons \text{Illite} \rightleftharpoons \text{Vermiculite} \rightleftharpoons \text{Montmorillonite}$$

The equation shifts to the right, increasing the lattice expansion, by decreasing the concentration of potassium; but the addition of potassium shifts the reaction to the left.

3. THE SESQUIOXIDES

The term *sesquioxides* is used in pedology for the hydrated oxides (or hydroxides) of iron, aluminium, titanium and manganese. Iron and aluminium oxides are the most important members of the sesquioxides. In Chapter 2 it was pointed out that iron is one of the main colour pigments in soil, while aluminium oxides are white. Manganese and titanium oxides are often dull black in colour.

When soils are highly leached under aerobic conditions, silicon is depleted more rapidly than the sesquioxides, which accumulate in the soil. Because iron and aluminium are more abundant in rocks than manganese and titanium, they accumulate at a faster rate and exert a stronger influence on soil morphology. Iron and aluminium oxides range from amorphous to crystalline in their organisation.

The hydrated iron oxides have already been discussed under 'Soil Colour' in Chapter 2. They form a continuous series from the less hydrated, red types, such as haematite to the more hydrated, yellow types, such as geothite. Unhydrated, black iron oxides, such as magnetite, contain oxygen atoms in the closest packing system with the co-ordinating iron in tetrahedral and octahedral positions. All of the iron oxides, once formed from the cations released during weathering, have very low solubilities and may survive many cycles of weathering in the soil.

The hydrated aluminium oxides form a series similar to that of the hydrated iron oxides. The most common aluminium oxide in soils is gibbsite which consists of a double layer of hydroxyl groups enclosing aluminium in the octahedral position. Boehmite and diaspore contain octahedral aluminium co-ordinated by oxygen and hydroxyl. Bauxite, which is mined for aluminium, is mostly a mixture of gibbsite and boehmite (its reddish colour comes from iron 'impurities').

Rutile and anatase are common primary titanium oxides found in soils, but anatase may form authigenically in the soil. In some iron oxides, titanium may substitute for iron (e.g., titanomagnetite and titanomaghaemite). Amorphous, hydrated titanium oxides, such as leucoxene, are found in the sand and clay fractions of many soils.

Manganese rarely occurs in soils as pure oxides. It is more usually associated with other elements.

The sesquioxides are more resistant to weathering than any of the silicate clay minerals. They have the lowest cation exchange capacity, the smallest ability for shrinking and swelling. They are therefore the common constituents of the intensely weathered tropical soils in which silica leaching has occurred.

The type of clay mineral that forms in a soil depends on a number of factors. The first factor is the nature of the parent minerals: certain primary minerals can form only a limited variety of clay minerals. The breakdown of the parent minerals limits the available components for clay formation. The second and perhaps more important set of variables is the weathering environment. Under neutral or alkaline weathering conditions (in arid or semi-arid areas) there is an ample supply of cations and less

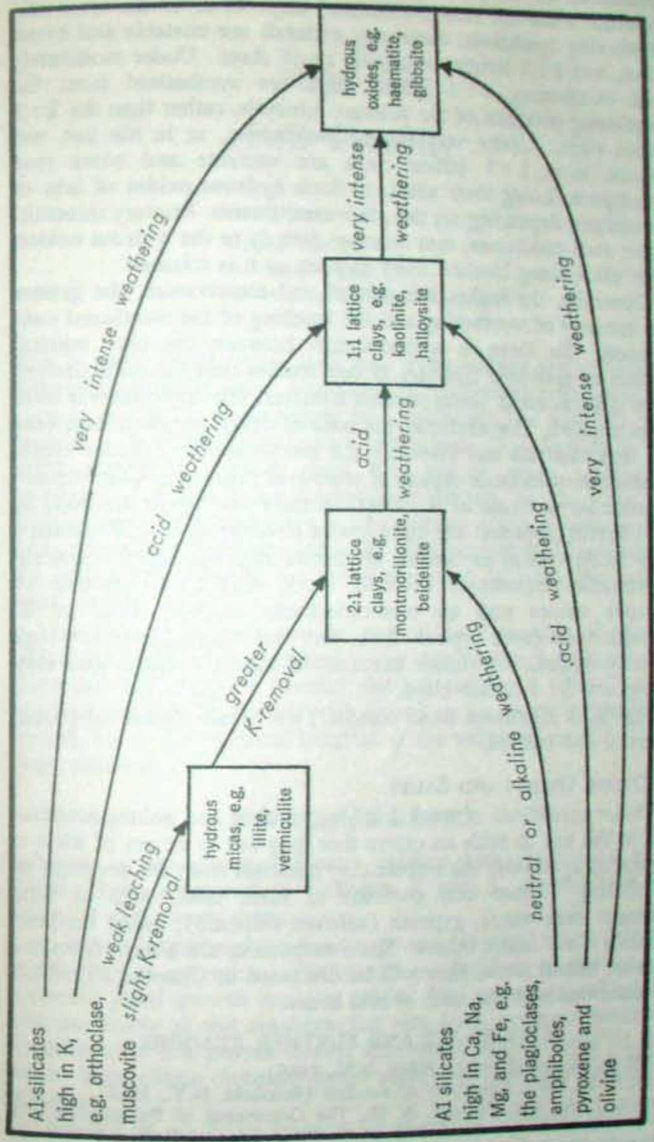

Figure 3.13: Influence of weathering conditions on clay mineral formation.

breakdown of the primary silicate structures, so the 2 : 1 lattice (hydrous mica and montmorillonite) clays form. Under more acid weathering conditions, these clay minerals are unstable and break down into 1 : 1 lattice (kaolinitic type) clays. Under moderately acid weathering, 1 : 1 lattice clays are synthesised from the weathering products of the primary minerals, rather than the 2 : 1 lattice clays. Under very intense weathering, as in the hot, wet tropics, even 1 : 1 lattice clays are unstable and some may decompose, losing their silica, to form hydrous oxides of iron or aluminium, depending on the other constituents. Primary minerals, under such conditions, may weather directly to the hydrous oxides, their silica being leached away as soon as it is released.

Generally, the higher the rainfall and temperature, the greater the intensity of weathering and the leaching of the weathered constituents. So there is a correlation between the clay mineral content of soils and climate. If one studies only the soils derived from igneous rocks under various climates, this correlation is even more marked. For example, the soils of desert areas contain very few clay minerals and these are the less stable 2 : 1 lattice types. Chernozem soils (soils typical of semi-arid regions) usually contain appreciable amounts of 2 : 1 lattice clays and lesser amounts of 1 : 1 lattice clays and are quite low in hydrous oxides. Podzolised soils (soils typical of humid, temperate regions) usually contain appreciable amounts of kaolinite clays, with lesser amounts of hydrous oxides and no montmorillonite or very little of it. Lateritic soils (soils typical of the hot, wet tropics) are very high in sesquioxides, with lesser amounts of kaolinitic clays and very rarely any 2 : 1 lattice clays.

Fig. 3.13 illustrates these trends. Two broad classes of parent material are treated.

4. OTHER OXIDES AND SALTS

Under conditions of weak leaching, cations and anions accumulate in the soil to such an extent that they are in excess of what is needed to synthesise the silicate clay minerals from the products of weathering. These ions combine to form salts, such as lime (calcium carbonate), gypsum (calcium sulphate), halite (sodium chloride) and many others. Since these salts are absent from the soils of humid areas, they will be discussed in Chapter 11, which is concerned with the soils of arid areas.

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Chapter Four

Soil Fertility

CONTENTS

FOR A VERY LONG TIME MAN HAS BEEN AWARE OF FERTILITY differences between soils. Variations in fertility are reflected in the natural vegetation even before man cultivates the soil, but they become more pronounced after cultivation. Soil fertility is the soil's ability to support plant life, and it poses world-wide problems, particularly for Agricultural Science. In this book soil fertility will be treated only in the present chapter because our main concern is pedology, but a list of references will be found at the end of this chapter for those who wish to study the matter more deeply.

Low fertility in soils can be caused by a number of factors which can be broadly classified under two headings, physical and chemical, depending on whether the low fertility is related to the physical features of the soil such as water, air space, structure or texture, or to the chemical features of the soil such as *pH*, or the plant nutrients.

PHYSICAL FACTORS

Many of the physical factors causing lack of fertility have already been discussed in Chapters 2 and 3, so they will be reviewed only briefly here.

In Australia perhaps the most important physical factor limiting soil fertility is soil water. The moisture conditions of the soil for optimum plant growth are just under field capacity, and, since soils are rarely at this condition for very long, optimum moisture conditions do not prevail during the whole growth cycle of the plant. Most plants therefore suffer some water stress during their lifetime.

The need of plants for water is more critical at certain stages in their growth than at others. For example, seeds will not germinate in very dry soils; for this reason, the conservation of water in the soils of the Australian wheat belt prior to sowing is vital. On the other hand, too much water affects some plants unfavourably at certain stages in their growth; thus heavy rain during the ripening of wheat is associated with rust and lower yields.

Plants differ in their tolerance of water stress. Hygrophytic plants are very intolerant of stress and require soils well above wilting point at all times, whereas xerophytic plants have mechanisms for withstanding dry conditions. Some xerophytic plants have low transpiration ratios and so take in less water (e.g., trees with small needle-like or waxy leaves). Others have in-built reservoirs which store water for use in dry seasons (e.g., the succulent flesh of cacti and the bole of the mallee tree). Tropophytic plants change their transpiration ratios in keeping with seasonal stresses; they are hygrophytic in the wet season and xerophytic in the dry season (e.g., the deciduous trees of monsoon woodlands). Thus the native vegetation adapts to the soil moisture conditions. But not all the crops man introduces into new areas are adapted to their new environments, so it is not surprising that low yields in some crops correlate with water stress during growth. Grain crops are now grown under a wide variety of climatic conditions in every continent, except Antarctica; some are subjected to water stresses far more severe than those that occur in their homeland.

Of course, man attempts by irrigation to offset losses of yield caused by water stress. But irrigation is usually not profitable or practicable on soils that have physical disadvantages in addition to a shortage of water. Very sandy soils, with their low water-holding capacity, lose irrigation water almost as soon as it is applied. Very clayey soils are difficult to irrigate because water penetrates them only slowly; but when it does eventually penetrate, they retain it well. Texture therefore greatly influences the soil's moisture status, producing striking variations in the vegetation of certain areas. In the Riverina irrigation areas, for example, fruit trees are grown on the sandier, well-drained soils of the ridges and levees, while rice crops which need a lot of water are grown on the clayey, poorly drained lowland soils.

Texture is also important in its effect on *tilth*, which is the cultivation of the soil to produce a loose, crumbly, friable, finely granulated state. Loams are the most suitable textures for cultivation as they have enough clay to form the fine granular structure and enough sand to keep them in a friable state. Sandy soils develop poor tilth because they lack binding agents, but clayey soils have poor tilth for the opposite reasons—they contain too many binding agents, and therefore tend to clod. Structure as well as texture is important to good tilth. Soils which have well-

developed peds are more easily broken into fine granules than are soils without peds.

Air capacity is just as important as moisture in soil fertility because yields sharply decrease as the soil's air capacity diminishes. Poorly structured soils and soils that are too clayey often lack aeration.

The soils most physically suited to cultivation are loamy textured, and well-structured, and contain sufficient organic matter to prevent breaking down of the structure after cultivation. An active microbial population is desirable. Other physical properties, such as self-mulching, also contribute to high fertility. Of course, not all soils possess these properties, so the farmer must do the best he can with the soil he has. The addition of lime, organic matter, a microbial population and, nowadays, even chemicals, helps to retain the soil in a suitable physical state for cultivation, but degeneration of the soil's physical properties after prolonged cultivation is still a major cause of decrease in fertility. Although textbooks tend to stress the chemical aspects of soil fertility, the physical aspects are certainly no less important.

CHEMICAL FACTORS

The chemical reasons for low soil fertility are the lack of the elements essential for plant growth. These elements are referred to as *plant nutrients*: they form what is called the dry matter of plants—that which remains after the water-content is removed. The amounts of these nutrients needed by plants may be calculated by analysing the dry matter (obtained by the oven-drying of plant tissue) in a laboratory. Plants vary in the quantities of the nutrients they require, and feed more heavily on some than on others at different stages in their growth. The nutrients essential for plant growth may be divided into:

1. *Tissue builders*. These are carbon, hydrogen and oxygen, and they are in plentiful supply in the air and water of the soil. There is therefore never a shortage of these elements unless the soil dries out.

2. *Major elements*. These are nitrogen, phosphorus, potassium and calcium. They are used in large amounts by plants, and after cropping they frequently need to be replaced in the soil.

3. *Minor elements*. These are used in lesser amounts by plants than are the major elements, but they are used in greater amounts than are the trace elements. Sulphur and magnesium may need replacing in the soil, but other minor elements, such as silicon and the sesquioxides (iron, aluminium, manganese and titanium), are the most common constituents of soils. These elements are there-

fore not toxic in large amounts in the soil, although they may interfere with the absorption of other elements.

4. *Trace elements.* These are needed by plants but only in minute quantities. Even in quite low concentrations they can be toxic to some plants. In fact, they are sometimes used as poisons to destroy weeds. Yet their absence can seriously retard plant growth. Examples of trace elements that are toxic in quite low concentrations are copper, boron, cobalt, zinc, vanadium and molybdenum. Plants need minute amounts of other elements which are not toxic at such low concentrations, for example, chlorine, fluorine and iodine.

1. THE TISSUE BUILDERS

Carbon, which is the major element of organic structures (just as silicon is in the inorganic world) is metabolised initially by plants in the form of carbon dioxide. Since the supply of carbon dioxide in the atmosphere is limited, it is essential that organic debris from plants and animals decompose to form carbon dioxide. Carbon dioxide is converted to organic carbon mainly through the photosynthetic activities of the higher plants and, to a lesser extent, by micro-organisms. Animals digest plants, converting them into tissue, and in the process they respire carbon dioxide. The whole process constitutes the carbon cycle which is concerned with the fixation and regeneration of carbon dioxide. More than 90% of the dry matter of plants is composed of carbon compounds of which cellulose, hemicellulose, lignin and starch are the main components. The decomposition of these substances into carbon dioxide involves the initial conversion of the long complex molecules into simple, easily metabolised derivatives.

2. THE MAJOR ELEMENTS

Millions of dollars are spent in artificially replacing these elements after cropping, because nature unaided, cannot replenish them as quickly as man removes them by harvesting crops or feeding crops to animals. Nowadays replacement by means of chemical fertilizer is popular, but there are other ways of making up the loss, and these will be considered as each element is discussed.

Nitrogen

The nitrogen content of the surface layer of most soils ranges from 0.02% to 0.4%; but from 92% to 96% of this nitrogen exists as insoluble organic combinations which cannot be extracted by plants. Thus although most soils contain enough nitrogen for plant needs, much of it is in a form that is quite useless to plants. All plants, with the exception of legumes (and a few less important groups), obtain their nitrogen directly from the soil solution as ammonium (NH_4^+) or nitrate (NO_3^-) ions. Legumes can extract some or all of their nitrogen requirements from gaseous nitrogen circulating through the soil.

The transformation of insoluble nitrogen into soluble nitrogen, the intake of nitrogen by plants, the consumption of plants by animals, and the return of nitrogen from plants and animals to the soil are shown in Fig. 4.1.

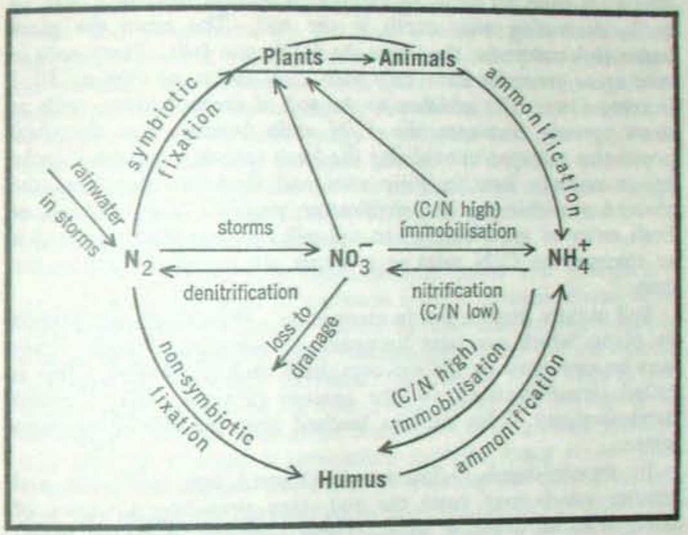

Figure 4.1: The nitrogen cycle.

When plants or their leaves die, micro-organisms attack the residue to form ammonia. This is termed *ammonification* and the ammonium ion thus formed soon attaches itself to the soil colloids and so does not exist in a free state, except in alkaline soils. In acid and neutral soils, ammonium ions are rapidly transformed into nitrate ions, especially if these soils are moist and well-aerated. This process is termed *nitrification.* Nitrate ions are anions, so unlike ammonium ions (which are cations) they are less likely to be adsorbed onto the negative surface of soil colloids. They exist freely in the soil solution and for this reason, they are easily leached from the soil. The rate of nitrification largely controls the quantity of nitrogen available to plants. Nitrification is further hindered by some of the soil's micro-organisms which compete with the nitrification process for the ammonium ions in the soil. The needs of these micro-organisms often leave a deficiency of ammonium ions so the nitrification process is slowed because of lack of the source material. When microbial activity is high, there is usually a serious shortage of ammonium ions for nitrification.

The rate of nitrification depends on the *carbon:nitrogen ratio* (C/N ratio) in the soil. Soil organic matter contains roughly 58% carbon and 5% nitrogen, giving a C/N ratio of 11.6 to 1. The C/N ratio for most agricultural soils lies between 9 and 14 to 1, decreasing with depth in the soil. The more the plant material decomposes, the lower the C/N ratio falls. Peaty soils in cold areas generally have very high C/N ratios, as high as 50:1 in some cases. The addition to the soil of carbohydrates, such as straw, greatly increases the C/N ratio because the microbial population engaged in oxidising the fresh carbon compounds locks up ammonium ions in their cells and therefore decreases the amount available for the nitrification process. The addition of fresh straw or grass cuttings to soil will result in crop failure if it so changes the C/N ratio as to cause nitrogen starvation in the crop.

Soil nitrates may be lost in many ways. They may be consumed by plants which are later harvested or eaten by animals. They may be converted by the soil organisms back to ammonia (this is called immobilisation) or into gaseous nitrogen (this is called denitrification). They may be leached from the soil by drainage water.

In thunderstorms, rainwater is charged with ammonia and nitrates which may enter the soil, thus providing a source of nitrogen in an available form. These additions, although appreciable over a year, would probably be no greater than the nitrate losses caused by leaching; so from the farmer's point of view they may not be very important.

Far more important to the farmer is the fixation of nitrogen by legumes. This is termed *symbiotic fixation*. The bacterium, rhizobium, in the nodules of legumes (in Australia, clovers and medicks are the most important) can extract nitrogen from the soil atmosphere and incorporate it into the host plant. The nitrogen that is fixed (i.e., removed from the soil and taken into the plant) is made available for other crops only when the host plant dies and is ploughed into the soil. This is what is done with legumes. A good crop of clover may fix 50 pounds of nitrogen per acre. About 20 pounds of nitrogen per acre might be used annually by wheat, and 18 pounds might be used on a dairy farm carrying one cow to two acres; thus the amount of nitrogen fixed by a clover crop together with additions from natural nitrification

could last for three years. (One can readily examine the nitrogen-fixers of legumes by locating the small nodules on the roots of clover, peas or beans.)

The symbiotic fixation of nitrogen is far more important than *non-symbiotic fixation* which is the fixation of atmospheric nitrogen by organisms which exist independently in the soil (unlike rhizobia which depend on the host plant). These organisms account for the better-than-expected nitrogen status of soils when losses of nitrogen for non-leguminous crops have been calculated. Azotobacter and Clostridium are the most widely studied independent nitrogen fixers, but other organisms such as the blue-green algae are probably also important. Non-symbiotic fixation is a growth-linked process, and the organisms require an appreciable amount of organic matter for their energy in order to fix a relatively small amount of atmospheric nitrogen. From 50 to 2,000 pounds of organic matter is required for every pound of nitrogen that is fixed. Because of these factors, biologists generally consider the effects of non-symbiotic fixation to be slight.

Since it is not always practicable to plant and plough-in a legume every few years to maintain a soil's nitrogen status, other measures must be taken to ensure an adequate nitrogen supply. The addition of suitable green manure, preferably from a legume, is beneficial. Chemical fertilizers (as distinct from organic fertilizers such as manure or blood and bone) are becoming increasingly favoured. One of the cheapest of the chemical fertilizers that provides nitrogen is ammonium sulphate; being an acid salt, it has the unfortunate side-effect of lowering the *pH* of the soil and this makes the nitrification in acid soils rather sluggish (the optimum *pH* for nitrification is between 6 and 7). The addition of ammonium sulphate without lime may therefore actually slow down the natural mechanisms for converting unavailable nitrogen into an available (ionic) form. However, ammonium sulphate is still the preferred means of chemically adding nitrogen to the soil in Australia. In America, preference is given to liquid ammonia which, injected into the soil under pressure, is held as ammonium ions. Chemical fertilizers have the advantage over organic fertilizers of providing a supply of immediately available nitrogen for the plant. Organic residues, such as stable manure mixed through straw, even though quite thoroughly decomposed when added to the soil, do not supply much nitrogen immediately, but they give a gradual supply over the next few years. Chemical nitrogen must be added with each new crop, for very little, if any, remains after the previous crop. Sugar cane is the only crop in Australia that is regularly given appreciable quantities of nitrogenous fertilizer; in most cases it is spread during sowing.

Phosphorus

Phosphorus is an essential constituent of every living cell and is especially important in reproduction and photosynthesis. It is found in the soil's organic fraction, but unlike nitrogen it is also found in inorganic compounds within the soil. Soil organic matter accounts for from 15% to 80% of the total phosphorus in the soil. Some primary minerals contain phosphorus (e.g., apatite, an important accessory mineral in igneous rocks). Phosphorus usually occurs in minerals in combination with calcium, magnesium, iron and aluminium.

Like all the other plant nutrients, phosphorus is available to plants only in the ionic form. Thus although a soil may contain appreciable phosphorus in both organic and inorganic combinations, there may be very little phosphorus available to plants. The availability of phosphorus poses a greater problem than that of any other nutrient because it is the most easily *fixated*. Available phosphorus may combine with other elements to form very insoluble compounds which plants cannot extract: this is termed 'the fixation of phosphorus'. Fixation can happen to added (fertilizer) as well as to native phosphorus.

The organic phosphorus fraction of soils is quite variable. The carbon:nitrogen:phosphorus ratio of soils may be close to 100:10:1. Fortunately, plants remove far less phosphorus from soils than they do nitrogen—a wheat crop might remove about one-quarter as much phosphorus as nitrogen.

The availability of phosphorus to plants in soils depends on the rate of formation of the three phosphate ions which plants can extract. These ions are:

$H_2PO_4^-$	(monovalent)
HPO_4^{2-}	(bivalent)
PO_4^{3-}	(trivalent)

Plants can extract the monovalent ion most readily. The concentration of the three ions in the soil solution is related to *pH*. At *pH* values less than 6.7, the monovalent ion dominates; between *pH* 6.7 and 8.0, although the bivalent dominates, the trivalent ion is also important. The availability of phosphorus depends on the conversion of organic and rock phosphates into these three ions. The reactions which bring this conversion about are slow and rarely keep pace with the removal of phosphorus from the soil by plants and animals, nor with the fixation of phosphorus.

Plants starved of phosphorus are retarded at every stage of their growth. Phosphorus deficiency in plants is not as easy to diagnose as nitrate deficiency. Plants can be severely starved of phosphorus

yet show no visible signs. By the time a deficiency is recognised it may be too late to remedy: plants take up much of their phosphorus in their early stages, and there is no hope of later additions making up for early starvation.

By world standards, Australian soils are particularly low in phosphorus: many contain less than 0.02% of total phosphorus, much of which is unavailable to plants. Phosphorus is the most widely deficient of all the plant nutrients in Australian soils, so it is not surprising that superphosphate, $Ca(H_2PO_4)_2$, is Australia's most popular fertilizer. Superphosphate is prepared from rock phosphate by the addition of sulphuric acid. Superphosphate is water-soluble, but triple-super ($CaHPO_4$) which is also widely used, is only slightly soluble in water but dissolves in neutral citrate solutions. Although superphosphate is soluble in water, it is not leached through the soil as one would expect: rather, it is precipitated close to the surface and accumulates in the soil over the years. Unlike nitrates, phosphates therefore are not subject to serious leaching from the soil except in very sandy soils that experience seasonal concentration of precipitation. The change from soluble phosphorus in superphosphate to insoluble phosphorus in the soil is termed *reversion*. This change occurs only in the soil, because superphosphate on its own does not revert. The reverted precipitate in the soil differs chemically from ordinary superphosphate. Although superphosphate itself has a *pH* of 4.5, soils seldom become acid after its addition (cf. sulphate of ammonia) because either the plant extracts the acid ion ($H_2PO_4^-$) as soon as it forms, or the ion is fixated.

Phosphorus fixation depends on the initial state in which the superphosphate is precipitated near the surface of the soil. If the *pH* of the soil is less than 7.5, it is most desirable that the phosphates be precipitated as calcium compounds, for calcium phosphate is sufficiently soluble to supply the plant's needs, but not soluble enough to suffer from leaching. If the *pH* is higher than 7.5 or if lime is present, hydroxyapatite may form from the phosphate; it has a very low solubility and releases insufficient phosphorus to supply plants. Unfortunately, in acid soils, the desirable calcium phosphates do not always form, and the phosphates complex with iron and aluminium to form very insoluble compounds; thus the phosphorus is fixated. Iron and aluminium oxides and kaolinite may all fixate phosphorus. Red soils of tropical and sub-tropical regions are particularly high in these compounds and have the worst reputation for fixating phosphorus. In krasnozems (red, well-structured soils) near Lismore, New South Wales, free ferric oxide exceeds 10% and, although the total phosphate in these soils is high, there is a serious phosphate deficiency in crops. Any added superphosphate is soon fixated. The addition of small amounts of lime with the superphosphate helps to reverse the fixation reaction and to maintain the *pH* of the soil between 6.0 and 6.5, the optimum *pH* for phosphorus availability.

The amounts of native, total and available phosphorus diminish with depth in most soils, unless there are a lot of phosphate-bearing minerals in the parent material, when the native phosphorus may increase with depth. The amount of available phosphorus diminishes even more than total phosphorus in illuviated soils, since fixation is greatest in heavier soils (i.e., soils with more colloidal components). One would expect available phosphorus to be lower at the surface where plant roots extract it.

Superphosphate benefits all crops in Australia, and it is also widely applied to pastures, sometimes by aerial top-dressing, an increasingly favoured method. Repeated application of phosphatic fertilizer causes a build-up of phosphates in the soil, so perhaps, over the years, Australian soils may not have such a low phosphate status in comparison with European soils which have been cultivated for a much longer period.

Potassium

Nitrogen, phosphorus and potassium are the three elements most likely to be found to be deficient when crop yields fall. Unlike nitrogen and phosphorus, potassium is not derived mainly from the soil's organic complex, but, like calcium, is a common constituent of many minerals such as the feldspars. Plants need it in quite large quantities; indeed, in quantities comparable to their need for nitrogen. The soils of dry and sub-humid climates usually contain sufficient available potassium for plant needs. This applies to the whole of the Australian wheat belt where, so far, no potassium deficiences have been reported. However, serious deficiencies may occur in wet areas, especially in sandy soils.

Potassium is important in the photosynthesis processes and in the formation of amino acids and proteins from ammonium ions. A soil's supply of potassium may be adequate in conditions where nitrogen and phosphorus supplies are low, but if these are then added as fertilizer a potassium deficiency may follow, and this deficiency is often shown by the premature death of plant leaves. The addition of potassium in some cases not only increases plant yields, but makes plants more resistant to physical stresses; for example, it may increase the winter hardiness of lucerne by prompting it to store more carbohydrate and protein.

Potassium, like phosphorus, may become fixated and hence unavailable to plants. Formerly, fixation was believed to occur only on drying, but now it is known to occur also in permanently moist soils. The effects of this fixation are not as serious as for phosphorus, because the available form of potassium is a cation (K^+) which features permanently in exchange reactions on the soil colloids (anionic exchange is known to occur, but it is less important than cationic exchange; see Chapter 3). However, potassium is far more easily leached from the soil than phosphorus.

When potassium is applied as fertilizer (e.g., as muriate of potash, KCl, or potassium carbonate, K_2CO_3), most of it is adsorbed onto the soil colloids and the part that remains in solution is usually leached, so that very little remains immediately available for plants. The optimum *pH* for potassium availability is between 6 and 8, which fortunately coincides with the optimum *pH* for nitrogen and phosphorus availabilities.

Some silicate clays contain potassium in their structures (e.g., montmorillonite and illite), so soils high in these clays, such as the chernozems, usually do not have a potassium deficiency. But the kaolinitic clays do not have room for potassium in their structures, so soils high in these and in the hydrated oxides (e.g., haematite and gibbsite) are often deficient in potassium; for instance the reddish tropical soils are thus deficient.

Calcium

It was pointed out in Chapter 2 that calcium is added to soil primarily as a *pH* corrective to maintain soil *pH* between 6 and 7, which coincides with optimum nitrate and phosphate availability. Liming a soil may in this way rectify deficiencies in these other elements. Unlike phosphorus, calcium does not suffer from fixation reactions. It is available to plants as the simple cation (Ca^{2+}).

Lime is usually added to the soil as ground limestone or lime. The term *lime* is used for all calcium compounds added to the soil. The main compounds are calcium oxide, calcium hydroxide and calcium carbonate. All these have the same long-term effect, the oxide changing into the hydroxide, which in turn changes into carbonate. However, the oxide and hydroxide raise the *pH* much higher and much more rapidly than does the carbonate, which is usually preferred because of its cheapness unless the *pH* situation is a very serious one.

The main sources of carbonate lime are calcite ($CaCO_3$) and dolomite, $CaMg(CO_3)_2$, which are both sparingly soluble. Water, containing dissolved carbon dioxide, reacts slowly with these to form calcium bicarbonate, $Ca(HCO_3)_2$, which, being soluble, allows calcium ions to enter the soil's exchangeable cations to replace some of the hydrogen on the colloids. The rate of this reaction depends on the amount of surface area of the lime that is exposed and the concentration of bicarbonate in the soil solution near the lime. Once this solution becomes saturated, it must diffuse away and be replaced by the unsaturated solution before more lime dissolves. Since carbon dioxide is essential to the reaction, the higher its concentration, the faster the solution near the lime. Carbon dioxide content is higher in wet soils because the air content, and hence the rate of renewal of soil air with atmospheric air, is lower. Carbon dioxide content is also higher in warm, acid soils. Fortunately, as soil neutrality is approached, the reaction slows down: it is undesirable to over-lime soils.

Over-liming reduces the availability of nitrogen and phosphorus and may cause a serious shortage of iron and manganese, which are not soluble at high *pH* values. Over-limed soils may be acidified by the addition of sulphur or aluminium sulphate.

Calcium, as well as correcting the soil's *pH*, is needed in quite large quantities by plants. Lucerne may contain 3% of calcium. The amount of calcium in a soil is usually highest near the surface where plants return it to the soil. Calcium deficiencies in plants cause stunting which is apparent in both roots and leaves, making diagnosis, assisted by *pH* measurements, relatively easy. Calcium is a common constituent of many minerals (e.g., the feldspars) and is far more important in rocks than phosphorus.

3. THE MINOR ELEMENTS

Sulphur is not required in large quantities by growing plants but is nevertheless an important nutrient. Its cycle in nature is similar to that of nitrogen. Sulphur compounds undergo many transformations as a result of the activities of plants, animals and microbes. Most of the soil's supply of sulphur lies in its organic matter, but a small amount is derived from minerals such as pyrites (iron sulphide). Plants assimilate sulphur from the soil in the anionic form (SO_4^{2-}), but small amounts of sulphur dioxide may be obtained from the sulphur dioxide in the atmosphere. Sulphur is returned to the soil by dying plants and by animals, and some may be lost by drainage (*cf.* nitrogen). The carbon:sulphur (C/S) ratio appears to be as important in the formation of available sulphates as the C/N ratio is in the formation of available nitrates.

The sulphur requirements of a soil are often fulfilled when superphosphate is applied, since superphosphate is prepared by the action of sulphuric acid on rock phosphate and contains a considerable amount of calcium sulphate. In fact, one hundredweight of superphosphate usually contains about 12 pounds of sulphur. Sulphate of ammonia, $(NH_4)_2SO_4$, which is usually

applied to correct nitrogen rather than sulphur deficiency, also corrects sulphur deficiencies. Sulphur may be used as a *pH* corrective when a soil is too acid.

Magnesium is needed by all green plants because it is a constituent of chlorophyll. It is also important in assisting the transportation of phosphorus in plants, so that the phosphorus content of a crop can be increased by adding a magnesium fertilizer. Magnesium silicates such as finely ground olivine or serpentine may be added to superphosphate to increase its effectiveness. Magnesium deficiency may be induced by unbalanced manuring or the repeated use of grass mulches.

Magnesium deficiency, uncommon in Australia, is often apparent in sandy soils which carry leafy crops such as tobacco. These soils are usually also deficient in calcium, so dolomite, $CaMg(SO_4)_2$, is often used to correct both deficiencies. By world standards, Australian soils are particularly high in magnesium. The belief that excess magnesium is harmful appears to be unfounded; the infertility of soils high in magnesium may be due to the presence of excess trace elements such as chromium, rather than to the magnesium.

Although many textbooks claim that plants require *sodium* in small amounts, it has not yet been demonstrated conclusively that this is so, perhaps because most soils contain some sodium brought across the continents as cyclic salt in rainwater (see Chapter 11). Problems associated with sodium excess rather than sodium deficiency are the main concern of soil nutritionists, especially in drier countries like Australia. It is not so much the sodium that is harmful to plants, but the anions with which it is associated. Sodium is commonly associated with chlorine as sodium chloride or common salt, and excess chlorine upsets plant metabolism. Sodium with sulphate or bicarbonate is not so harmful. Some crops are very intolerant of quite small amounts of sodium chloride—as little as 0.1% may lower the yield from citrus trees. Vines and some grasses tolerate much more sodium chloride. Some Australian soils in areas with arid or semi-arid climate are too saline for agriculture, and some are even too saline for reclamation by means of irrigation together with addition of gypsum to replace the sodium adsorbed onto the soil colloids. Irrigation using water containing more than 0.1% salt may aggravate the situation.

Silicon is unlikely to be deficient in soils, for it is found in most primary and secondary minerals. Plants appear to absorb silica as silicic acid (H_4SiO_4). They use quite large amounts of silicon just as they absorb large quantities of nitrogen. Too much silicon cannot harm plants.

All plants and animals require *iron*. Unfortunately, the iron in soil may be very insoluble and thus unavailable to plants. Even if the soil shows a red or yellow colour, this does not mean that there is a heavy concentration of iron, for strong colouring can be produced by quite small quantities of iron, and the colour-producing pigments are extremely stable in the soil. Iron is least available when *pH* values are high and when lime is present. Then some iron-sensitive plants such as vines fail to produce their normal supply of chlorophyll and so develop pale, chloretic leaves. In such cases iron is often applied to the soil as chelates, because the artificial acidification of the soil which would rectify the iron deficiency is too costly and would upset the availability of the other nutrients.

In acid soils, iron is more freely available to plants and iron deficiencies have not been reported from such soils. *Aluminium* is similarly available in acid soils, though it may occur in toxic proportions and interfere with the intake of phosphorus. So far no aluminium deficiencies have been reported in Australia.

Plants are able to adjust their intakes of *manganese* to a certain extent to suit the soil. In acid soils, manganese, like aluminium and iron, is freely available to plants and they may take in as much as 1,000 parts per million; but in alkaline soils, a deficiency may occur and the same species may take in only 10 parts per million. In Australia, manganese deficiency has caused crop failure, especially of oats, in soils with *pH* higher than 6.7, when bacteria convert available manganese into insoluble oxides. The easiest way to rectify a manganese deficiency is to acidify the soil with aluminium sulphate or sulphur, or to sterilise the soil by removing the bacteria. At low *pH* values, manganese may be available in toxic proportions.

Plants also require *titanium*, a common constituent of many ferromagnesian minerals. Australian soils have not, so far, been reported to be deficient in titanium.

4. The Trace Elements

The term *trace element* may be used for all elements required by plants in small amounts, being beneficial to plants in low concentrations of less than 1,000 parts per million, but becoming noxious above this level. Some researchers (e.g., Leeper, 1964) describe as trace elements all elements needed in very small quantities by the plant (including some of those discussed above as minor elements) which are not normally toxic within certain limits.

Trace elements known to be deficient in soils are copper (Cu), cobalt (Co), molybdenum (Mo), zinc (Zn), selenium (Se), boron (B), fluorine (F) and iodine (I), and also possibly vanadium (V) and chlorine (Cl). Others will undoubtedly be added to this list, possibly germanium (Ge), gold (Au), silver (Ag) and mercury (Hg), but very little is known at present about their behaviour in soils.

Since trace elements are required by the plant in very small quantities, it is difficult to see why some soils do not contain them. An acre of oats which may remove 7 pounds of phosphorus, may remove only ¼ ounce of copper, 1 to 2 ounces of zinc, and 7 ounces of manganese. Trace elements are usually concentrated by plants in the surface soil and decrease markedly with depth. Erosion of the topsoil could result in a serious loss of a soil's trace elements; indeed, many deficiencies are known to have arisen in this way.

The trace element content of a soil depends on the nature of the parent rock. Unfortunately, the trace elements are not evenly distributed throughout the various rocks; for example, cobalt and chromium are concentrated in basic rocks. The minerals plagioclase, pyroxene and olivine account for most of the trace elements found in soils developed from igneous rocks. Metamorphic rocks tend to be richer in trace elements because when rocks containing colloids are metamorphosed any adsorbed cations, some of which are trace elements, are incorporated into the new rock. The distribution of trace elements in a soil's parent rock enables one to anticipate with reasonable accuracy the possible trace element status of the soil. Difficulties arise in parent materials that have undergone multi-cyclic weathering.

Trace elements may be held in soils in the silicate clays, the sesquioxides and the organic colloids. Like the other nutrients, they may be adsorbed onto these colloids. A shortage of trace elements in the soil usually visibly affects the plant—the leaves may become bronzed or mottled, the growing points wither, or the growth habits alter. Some plants unfortunately do not indicate their trace element deficiency, so an analysis of plant tissue or a response to added trace elements may be the only means of diagnosing the deficiency. Because most trace elements have low toxic levels, any addition of such elements to the soil must be done with caution.

Plants vary in their intake of trace elements. Intake depends on the plants' needs as well as on soil conditions such as *pH*, moisture, nutrient status and the level of the particular element in the soil; for example, at low soil-copper levels, grasses contain more copper than clovers, but at high soil-copper levels, the position is reversed. Competition among plants is *therefore* *important*, as some plants are better able to cope with low trace element levels.

Trace elements enter into the general processes of soil formation. They may, for example, be concentrated in illuvial horizons with humus, silicate clays and sesquioxides. They are also susceptible to leaching.

A trace element may be added directly to a soil in which it is deficient; for example, powdered borax alone may be added. More usually, trace elements are added with other fertilizers. Many mixed fertilizers contain trace elements, as only a few ounces or pounds of them are required per acre. They may be sprayed directly onto the leaves of plants, or they may be inserted in pellet form under the bark of trees. Trace element deficiencies do not only arise from lack of the element in the soil. Like other nutrients, they may be present in the soil but be temporarily unavailable, so correction of the conditions causing their unavailability could eliminate the deficiency. A slight increase in acidity will release most trace elements, but not molybdenum; deficiency of this element can be overcome by placing sulphur near the plant's roots. Plants may suffer from deficiencies in more than one trace element at the same time, so several may have to be added together.

Sandy soils and soils high in organic matter may be deficient in *copper*. Cattle become unhealthy when grazed on pastures lacking copper; sheep, although healthy themselves, grow inferior wool. Deficiencies of copper and zinc produce interacting effects: plants may respond to zinc if copper also is given, but not if it is withheld. *Zinc* deficiencies are especially related to low *pH* values. Sunny climates are most likely to have zinc deficiencies, especially if the plant has a weak root system. Vines and fruit trees in the Victorian Mallee, a sunny area, improve greatly after spraying with zinc sulphate.

Boron deficiency affects the growing points of plants. Borax (usually 10 - 12 pounds per acre) must not be mixed with seeds when added to the soil, as it is poisonous until it ionises in the soil at lower concentrations. Sea water contains boron, marine sediments usually contain ample boron, so much so that soils derived from some sedimentary rocks suffer from boron toxicity. Boron deficiency may be induced by over-liming soils.

Molybdenum occurs in soils as the molybdate ion (MoO_4^{2-}) which silicate clays and sesquioxides can bind into an unavailable form under acid conditions. In neutral or alkaline soils, plants may take up too much molybdenum and animals may suffer from the excess. Its main advantage lies in the role it plays in nitrogen fixation as rhizobia, and the non-symbiotic organisms depend on

it to operate. Molybdenum is therefore often required to establish the legume, subterranean clover. A single application usually lasts for years.

Cobalt is also concerned with nitrogen fixation and is therefore essential for legumes. It is essential, too, for animals in that it helps to produce vitamin B_{12}.

Chronic poisoning in animals can result from grazing on pastures containing excessive *selenium*; on the other hand, various disorders, including muscular distrophy, can result from selenium deficiency. The need for the right balance of selenium emphasises that the application of trace elements is a very delicate procedure. Fortunately, a single application of most trace elements can rectify a deficiency for many years.

SUMMARY OF CHEMICAL FACTORS

Hydroponics clearly shows the plant's need for the major and some of the minor nutrients; if the culture solution lacks any of these nutrients, the plant will not grow. However, the need for trace elements is not so easily demonstrated, as such small amounts of them are required by plants that culture solutions made from distilled water and analytical reagents contain ample supplies as impurities. Their need is shown only by work with absolutely pure materials. Deficiencies in any of the nutrients result in lower yields, but there are often other means of detecting the deficiencies. The leaves of some plants may become discoloured when elements are deficient; for example, nitrogen deficiency produces yellowish leaves, and potassium deficiency a brown edge on the leaves.

All of the essential chemical elements enter the plants through their roots and pass into the stems, leaves, flowers and seeds. They return later to the soil as a result of the decomposition of the plants after they die, or the decomposition of the excreta and bodies of animals, many of which feed on the plants. Once at the surface of the soil, the nutrients produced by this decomposition may become insoluble and hence unavailable to plants, or they may be taken up again into the cycle of life, death and decomposition.

Whilst *pH* is clearly of vital importance in the availability of nutrients to plants, the student of soils must realise too that he cannot achieve an ideal blend of nutrients in a given soil, if by 'ideal' is meant a blend in which each nutrient preserves the optimal availability it is said to possess when it is considered in isolation; for example, the *pH* for optimal availability of nitrogen and phosphorus (between 6 and 7) may cause a serious shortage of some of the minor elements and trace elements. The farmer must always compromise, watching especially the various deficiencies that are most seriously affecting his yields. The nature of many deficiencies is sometimes difficult to recognise, for similar effects can be produced by several elements, so a soil or plant analysis is the only satisfactory means. Even so, a soil analysis of total elements may not indicate a deficiency as the problem is likely to be one of availability rather than total lack. The addition of certain elements can render others insoluble; conversely, it might free others for extraction; so the application of fertilizers has sometimes produced different results from those expected.

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Chapter Five

Soil-forming Factors

CONTENTS

ALL VARIATIONS IN SOIL PROPERTIES CAN BE ACCOUNTED FOR BY five factors: climate, parent material, organisms, relief, and time. *Climate* includes all atmospheric forces acting on the soil. *Parent material* is the material from which the soil horizons have developed; it includes pre-weathered materials such as alluvium, colluvium or loess, as well as fresh rock. *Organisms*, all the living forms or remains of living forms in the soil, includes vegetation as well as soil fauna and microfauna. *Relief*, the geomorphic features of the landscape, is the configuration of the soil surface including slope, exposure, position on the slope, and changes in slope away from the soil; the position of the water table within or beneath the profile may also be included as it is an important feature of the geomorphic environment. *Time* refers to the duration of weathering.

RELATING THE FACTORS

Hans Jenny, in 1941, regarded these five factors as independent variables and related them in a simple equation:

$$S = f\ (cl, o, p, r, t)$$

where S is a property of the soil, cl is the climate, o is the organisms, p is the parent material, r is the relief, t is the time or duration of weathering.

But consideration of them as independent variables is misleading because the similarity between maps of climatic, vegetation and soil regions suggests that at least some of the factors are intimately related. It would be more realistic to regard some of the factors as dependent variables, particularly climate and organisms. Parent material is a more independent variable, but even it is sometimes related to the other factors; for example, loess is commonly found on the lee side of deserts. Parent material may be related also to relief; for example, on slopes of more than three degrees, the soil parent material is more likely to be colluvium than fresh rock. When soils are developed from the underlying rock, parent material may be quite independent of relief.

Jenny later (1961) emphasised the dominance of any one factor in the development of certain soil profiles: the other factors, although still varying, were thus treated as subordinate. It is not difficult in the field to discover which one of the factors accounts for most of the variation between two differing profiles. The following equations (after Jenny, 1961') relate soil properties to the dominant factor of soil formation:

$S = f\ (cl, \mathrm{o, r, p, t})$	climofunction
$S = f\ (o, \mathrm{cl, r, p, t})$	biofunction
$S = f\ (p, \mathrm{cl, o, r, t})$	geofunction
$S = f\ (r, \mathrm{cl, p, o, t})$	topofunction
$S = f\ (t, \mathrm{cl, o, r, p})$	chronofunction

The effects of variations in any of these factors on soil properties may be assessed by studying soils in which four of the factors are as similar as possible while the fifth varies. Since some of the factors may be dependent, this assessment can be quite difficult, so the study of the dominant factor is seldom as simple as these equations imply.

Another serious objection to the factorial method of studying soil development is that it implies that soils have developed under constantly uniform conditions since their initial state. This may be true for very young soils which have not yet attained full horizon development, but it would not apply to soils thousands of years old, as climate has fluctuated quite noticeably in the immediate past and very markedly during the Pleistocene Period. Buried soils, quite different in morphology from their overlying soils, are commonly encountered in the field. Jenny's equations in no way accommodate buried soils. Stephens (1947) attempted to overcome the limitations of the factorial method by suggesting that the equations be changed to integrals. He wrote the main equation as

$$S = \int_{t\,=\,\text{B.P. } x \text{ yrs}}^{t\,=\,0} f\ (\mathrm{cl, o, r, p})\ dT$$

This equation indicates that climate, organisms, parent material and relief vary over time and that a soil is the integral of all the separate combinations of these factors at the point in time when it is examined. This equation makes allowance for the changes that take place in profile morphology as the environment changes. It also accounts for deep, complex soil sections in which remnants of buried soils are overlain by the present-day profile. The integral approach to the factors of soil formation is the most useful, but its complexity is such that we shall leave it until later in this chapter, turning now to consideration of the factors of soil formation in isolation, and reminding the reader that in reality the factors operate together, not in isolation.

In the ensuing discussion, it will be assumed that the factors other than the one being considered are controlled; for example, when relief is discussed, it will be assumed that the soils examined have similar climatic and biotic environments, are derived from similar parent materials, and have been weathering for similar durations, without change in the environment.

CLIMATE

Climate was one of the first factors of soil formation to be stressed in pedology; in fact, the early Russian soil classifications were based on it. Small differences in climate can cause marked differences in the morphology of soils. Because the climates of the world are gradational, with some exceptions, one would expect the main soil types to grade one into the other, but this is not

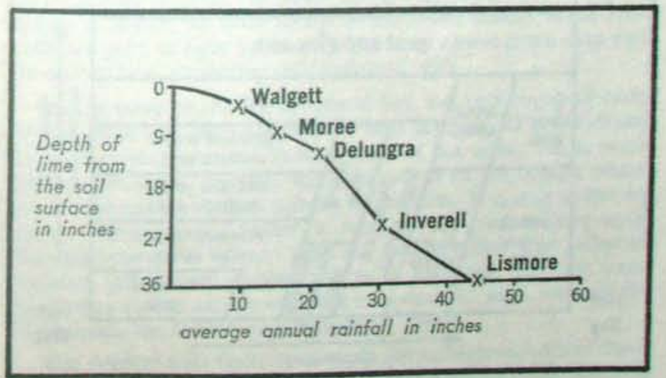

Figure 5.1: Relation between rainfall and depth of lime in basaltic soils.

always so, for the other factors of soil formation frequently intervene and give rise to sharp boundaries between soils. Sometimes, however, soils do form a continuum, and it is only after sufficient change has occurred that a different name is given to the soil. Such gradations are found in the basaltic soils of northern New South Wales, from the coast inland. The soils of the wetter coastal regions are red in colour; this colour is progressively lost as soils draw closer to the dry regions, grading through chocolate and brown into prairie soils in which the black humus masks the red colour. In the drier areas, lime begins to appear low down in the profiles and gradually becomes shallower further west through the black earths. The graph in Fig. 5.1 for middle slope soils on basalt indicates the relationship between the depth of the lime horizon from the surface and the average annual rainfall in northern New South Wales.

Finally, in very dry areas, heavy textured parent materials form soils which lack the red and even the black colourings of the soils further east. These soils are the grey and brown soils of heavy texture which usually contain lime and salt throughout the profile. The Great Soil Group names given to these soils therefore represent points along a continuum.

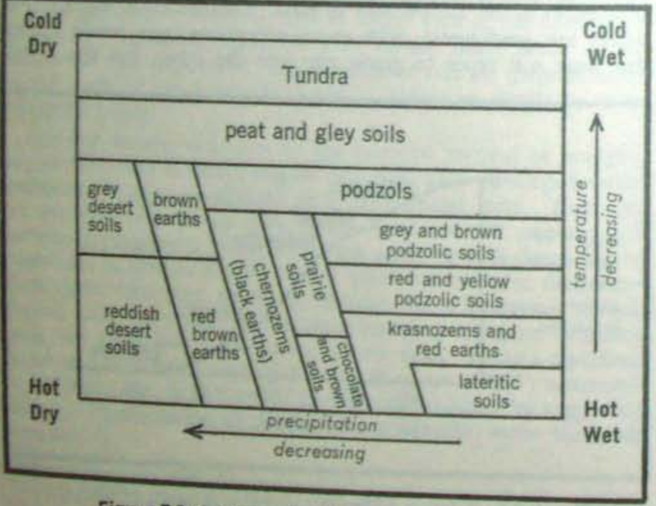

Figure 5.2: Soil zonation in response to climate.

The *zonal distribution* of soils which will be considered in more detail in later chapters relates Great Soil Groups to climate. A simplified zonation, in which both temperature and rainfall decrease, appears in Fig. 5.2. This zonation is only a model for the sake of classification, and it must be noted that the other factors of soil formation have not been included. There is, however, some correlation between this zonation and Great Soil Group maps of the continents, especially North America. One of the reasons for the low correlation is that temperature and rainfall do not vary uniformly across a continent as they do in the zonation model.

An examination of the effects of climate on individual soil properties will show how this zonation has arisen.

It was pointed out in Chapter 2 that the main *colour pigments* in soil are organic matter (black and brown), iron (red and yellow), and lime, soluble salts, unstained quartz and pure clays (white). Under very hot, wet conditions, organic matter decomposes more rapidly than it accumulates and tends to form the mobile, fulvic acid components which are not coloured. The lateritic soils and krasnozems are therefore lacking in black and brown colour pigments. Under wet, cooler conditions, fulvic acid forms and assists in the mobilisation of other constituents in the profile, but some humic acid accumulates in the illuvial horizons, so podzolised soils commonly contain black or brown *B* horizons. Under drier conditions, organic matter accumulates throughout the profile, especially in the *A* horizon, where humic acid, the less mobile and more coloured constituent of organic matter, imparts an intense black or dark brown colour. Under very cold conditions, organic matter decomposes less rapidly than it accumulates, so peats tend to form. In desert areas, there is insufficient organic matter to colour the soils, except in the cooler deserts where some soils are grey or light brown. However, in desert areas, salts tend to accumulate, producing pale coloured soils.

Iron is more prominent in soils of hot, wet regions where many of the other soil constituents have been leached. It is also found in appreciable quantities in most soils of hot areas, but in semi-arid to sub-humid climates the black colour of the organic matter masks the red and yellow colours of the iron. It is only in the dry areas where organic matter is minimal that red colours again become prominent. Desert soils are reddish rather than yellowish because yellow iron oxides are more hydrated than red ones. Since iron oxides are translocated in podzolised soils, some of the *B* horizons are reddish or yellowish in colour.

The deepest soils form under hot, wet conditions, so soil *depth* progressively decreases with both decreasing rainfall and temperature. Monocyclic desert and tundra soils are usually very shallow. The soils of temperate, humid and sub-humid regions tend to be intermediate in depth between the deeply weathered tropical soils and the shallow desert and tundra soils.

The higher the temperature and the greater the rainfall, the more rapidly do primary minerals decompose into secondary minerals; in other words, heat and high rainfall cause rapid formation of *clay* minerals. Tropical soils therefore tend to be high in colloids, especially the hydrated oxides, while desert soils tend to be sandy. Jenny, 1941, has drawn the graphs (Figs. 5.3 and 5.4) relating the percentage colloid in soils to both temperature and rainfall:

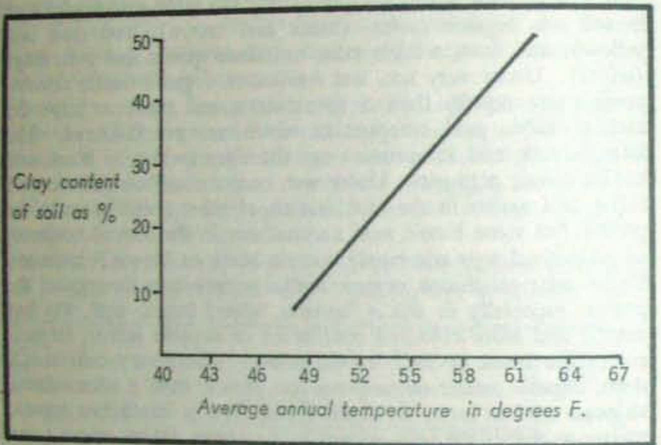

Figure 5.3: Clay content of soils derived from basic rocks with changing average temperatures. Samples taken in North America from 40 inches depth under the same average annual rainfall.

Most lateritic soils are clayey in *texture*, as some of the silica, which forms the greater part of the sand fraction of soils, is leached. The primary minerals in lateritic soils rapidly weather to release cations, anions and small crystal units which join together to form clay minerals. Even the very resistant minerals, such as quartz, are vulnerable under very hot, wet conditions. Tropical soils contain so much colloidal component that there is very little opportunity for clay illuviation, so texture differentiation is generally lacking. In podzolised soils, where colloid formation is not so rapid, but rainfall is quite high, clay illuviation is usually apparent and marked texture changes occur down the profile. In desert soils, there is usually not enough rainfall to bring about the illuviation of clay minerals, so again there is little texture differentiation.

The nature of soil colloids is also dependent on climate, as clay minerals are synthesised from the components which are not leached from the soil. Under tropical conditions, most of the soil's mineral matter is susceptible to leaching, but the sesquioxides accumulate to dominate the soil's colloidal fraction. Under moist, temperate conditions, most of the bases are leached from the soil, but silica accumulates with the sesquioxides, so 1:1 silicate clays and hydrated oxides form. Under semi-arid conditions, bases form an integral part of the silicate clay lattice and 2:1 lattice clays (e.g., montmorillonite which contains potassium in its structure) feature in the soil. In deserts, oxides and carbonates form an appreciable proportion of the soil's colloidal material because the bases are not all removed from the soil and so combine with anions.

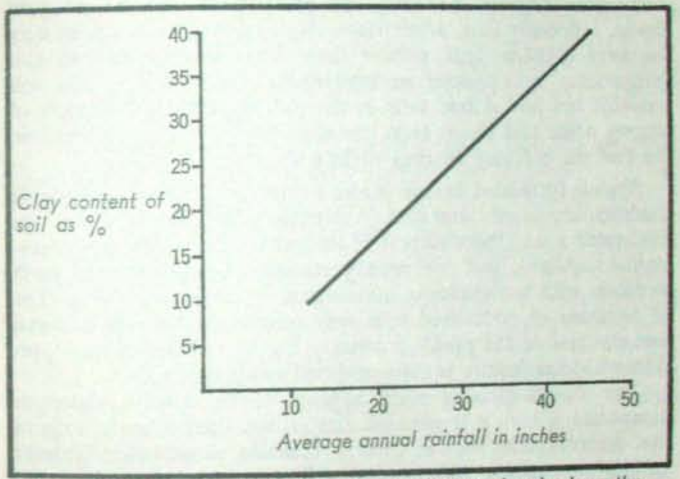

Figure 5.4: Clay content of soils (0-10 inches deep) along the 52° F. average annual isotherm in North America.

Since soil *structure* and the presence of peds depends upon the amount and nature of the soil colloids, structure is thus also related to climate. Optimum aggregate formation occurs in the presence of either 2:1 lattice clays and organic matter, or appreciable quantities of iron oxides. The best structured soils, therefore, are those of the semi-arid to sub-humid areas and the tropics. Desert soils generally have insufficient colloid for marked ped formation while podzolised soils usually have poorly structured *A* horizons, which are depleted of clay and sesquioxides, and better structured *B* horizons, which are enriched in clay and sesquioxides.

Soil *reaction* (*pH*) is related to both rainfall and temperature. Podzolised soils are very acid because the slow decomposition of the humus formed from forest litter allows humic acids to accumulate in the soil. Lateritic soils, on the other hand, are not so acid because the rapid decomposition destroys any humic acids that may form and so the soil *pH* is not greatly lowered. Bases are leached from both podzolised and laterised soils. Soils of sub-humid regions, however, are often quite alkaline, because bases, especially lime, accumulate. Soils of semi-arid regions may be very alkaline (*pH* greater than 8.5) because sodium and magnesium are present in appreciable quantities on the soil colloids but not as free salts in the soil. In deserts, free salts of strong acids and bases (e.g., sodium chloride) may accumulate, so that the *pH* may be only slightly above 7.

Nodule formation in soils is also related to climate. Sesquioxide nodules are most abundant in lateritic soils, especially in the indurated zone (the structure of all indurated zones is not necessarily nodular), and are usually scattered throughout red earth profiles, with tendencies to concentrate in certain horizons. The *B* horizons of podzolised soils may contain sesquioxide nodules, but the rest of the profile is usually low in nodules of any type. Lime nodules feature in semi-arid and sub-humid soils.

The mobilisation of silica appears to be closely related to temperature, for it is released and re-deposited in some soils of hot desert regions, such as Central Australia, as well as in the soils of hot-wet regions. Silica pans are therefore encountered in soils of hot-wet and hot-dry regions.

Any of the above generalisations may be disturbed by the inter-play of climate with any of the other factors of soil formation, so they represent trends which one would be unlikely to encounter in a transect across a continent unless the four other factors of soil formation were rigidly controlled, which is unlikely. The generalisations apply to virgin profiles because cultivation, especially with application of fertilizers, can cause marked changes in soil properties.

ORGANISMS

Because vegetation and climate are related, many of the above comments also apply here. However, the effect on soils of organisms, including vegetation, is rather different from the effect of climate. Climate is an external variable which may not change

during the development of a soil profile from fresh parent material. The organism factor, on the other hand, is an internal variable, for decaying vegetation and the microbial population are part of the soil's constituents. In addition to this difference, vegetation changes as a soil profile develops from fresh parent material. Climate is a more independent variable than vegetation; the only soil-forming factor that could condition it is relief. The organism factor is related not only to climate, but to parent material, relief and time as well, so it is a very dependent variable. It does, however, have some independent features; for example, the type of natural vegetation which develops in an area depends partly on seed access to that area, so that different types of vegetation may be found in regions of similar climate, parent material, relief and age in the various continents. Each continent's vegetation has evolved independently since the Jurassic when the continents drifted apart.

Soil formation and the development of vegetation are concomitant. When fresh parent material is first exposed to the forces of weathering and soil formation, the first plants colonise the surface. During the ensuing changes, vegetation and soil are intimately related and evolve together, increasing in complexity with age. It is impossible to say which came first, the soil or the vegetation, for they depend on each other. Many complicated changes occur after soil formation begins, before the soil becomes fully differentiated and the vegetation attains a constant species composition. These changes may extend over centuries and are referred to as *seral progression*. The end-product is termed the *climax* vegetation and the *mature* soil: the latter will be discussed in more detail later in this chapter.

When fresh rock is initially exposed to the forces of weathering, the first plants, if the area is cool and moist, are the lowly lichens, which attach themselves to the rocks by short outgrowths from beneath. When it rains, these plants take up water and renew their growth. They gradually etch away particles of rock and provide small amounts of organic matter to form a very thin, primitive soil in which a second community, the mosses, can become established. The mosses accelerate weathering and the production of organic matter so that a third community, the grasses together with annual and ephemeral plants, colonises the soil. Profile differentiation takes place and the soil becomes deeper and richer in organic matter and plant nutrients, enabling low scrub to be supported. If soil formation continues uninterrupted, a well-developed podzolised profile may form and support quite large trees, these representing the climax vegetation. Such a succession in middle latitudes may take centuries, but in tropical areas the succession takes far less time because weathering is more rapid.

Very rapid seral progression has occurred on the volcano Krakatau in the Sunda Strait between Java and Sumatra. The lower slopes of Krakatau were covered in tropical rainforest until 1883 when a violent eruption completely removed part of the island, and buried the rest beneath many feet of white-hot ash which sterilised the whole island of all organisms. Even if rainforest seeds had been available, they could not have germinated in the bare ash surface, so a complicated series of soils and vegetation had to develop before trees could grow.

During the first few years after the eruption, grasses established themselves in the ash, and by 1900 a cover of savanna had developed. The shallow soil that had formed could not hold much water, so drought-resisting shrubs gradually replaced the savanna and then drought-resistant trees moved in. By then, horizons had formed in the ash, and the surface few inches were darkened by organic matter The soil's nutrient status permitted invasion by tropical forest trees, so that today Krakatau is covered by vegetion similar to that before the eruption. However, the present flora is much poorer than the pre-1883 flora because a large number of species have not had sufficient time to cross the Sunda Strait. If the eruption had occurred in a less remote area, the flora would now undoubtedly be richer.

A soil property that exercises a most important influence on vegetation is water-holding capacity. This depends on the amount of colloid (both organic and inorganic) in the soil. The more fully developed the soil, the higher its colloid content and water-holding capacity, so that later plant communities are less drought-resistant than earlier ones. A different soil profile is associated with each vegetation succession, although climate, parent material and relief remain constant throughout the transformations. The factor *time*, of course, is changing with the vegetation and soil, so the most sensible way to consider seral progression is by the use of Stephens' integral equation (see the introduction to this Chapter) integrated with respect to time. In seral progression, parent material, relief and climate are constant and the soil and vegetation co-vary with time. More will be said about the evolution of soil profiles later in this Chapter.

Because of varying seed access, adjacent climax vegetations are not necessarily the same even if the other four factors of soil formation are similar. This is so in the sub-humid areas of the United States, on weathered moraine, where mature soils (the time factor is therefore controlled) are found under similar climates and relief. Deciduous hardwood forest is frequently

found side by side with prairie grassland. The soils under these two vegetations are very different. Weathering has apparently been more intense under forest as the soils contain less of the sand and gravel fractions and have distinct podzolic features. Clay and sesquioxides have been illuviated in the forest soil, but not in the prairie soil, which contains more organic matter and is blacker in colour. The humus in the forest soil presumably contains more of the mobile fulvic acid component and less of the humic acid, black-coloured component than the prairie soil. The carbon:nitrogen (C/N) ratio of the forest soil is higher than the prairie soil. Lime in the moraine is leached to a greater depth in the forest soil than in the prairie soil. The following graph indicates the changes in *pH* with depth in forest and prairie soils. Note that the *pH* is less variable in the prairie soil.

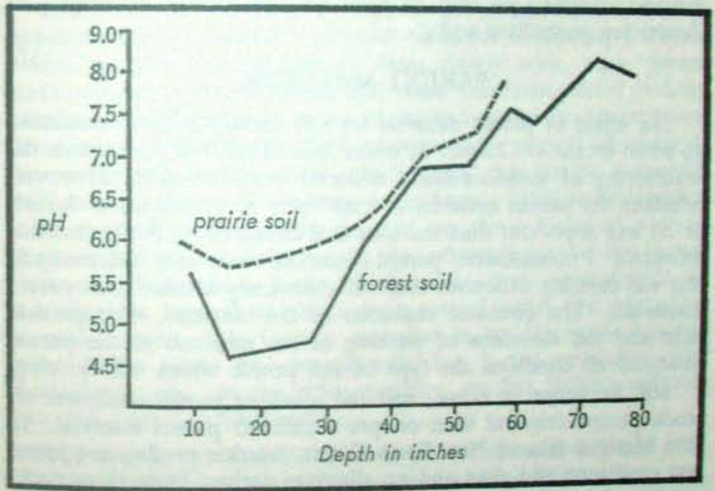

Figure 5.5: The pH of soils under prairie grassland and forest. The other soil forming factors are similar. *(Jenny, 1941.)*

These soil differences are probably traceable to vegetation; the deciduous forest stimulates leaching and accelerates soil profile development, especially base depletion and illuviation of mineral constituents, as forest organic matter decomposes more readily into mobile, less coloured constituents. Examples such as this are not so easy to find in Australia where grassland is associated with heavy textured clayey parent materials. In Northern-Central New South Wales, grassland with prairie soils or chernozems is found adjacent to woodland with podzolised soils, but the parent material, as well as the vegetation always differs, so that one cannot easily ascertain which properties are directly the result of vegetation differences. Also, it could be that the heavy texture of the parent material, rather than seed access, precludes the growth of trees in the grassland. The black colour of the grassland soils and illuviation of constituents in the woodland soils are definitely related to the vegetation.

The inherent differences between Australian and Northern Hemisphere podzolised soils may be attributable largely to the vegetation factor. The Australian eucalypt vegetation produces a leaf litter very different from that of the Northern Hemisphere deciduous, evergreen or coniferous forests. Eucalypt leaf litter decomposes quite rapidly (this could be the result of Australia's warmer climate) so that *O* horizons are poorly developed in Australian podzolised soils.

PARENT MATERIAL

The effect of parent material on soil formation was considered to some extent in Chapter 3, under Soil Mineral Matter, when the weathering of aluminosilicate minerals was discussed. However, whether the parent material is fresh rock or a weathered deposit is no less important than the chemical character of the constituent minerals. Pre-weathered parent materials react very differently to the soil-forming processes than do chemically similar rock parent materials. The chemical character of the minerals, their particle size and the closeness of packing of the minerals in the parent material all condition the type of soil profile which will develop.

Soil formation is slower and the resulting profile shallower on rock parent material than on pre-weathered parent material. In the Mudgee area of New South Wales, lateritic profiles are found on sandstone and slate and on alluvium derived from these rocks and from granite. Since the lateritic soils were formed during the Tertiary, only the mottled and pallid horizons remain in the Mudgee area today. On rock, these two horizons together range in thickness from 5 to 12 feet, but on alluvium they may total from 40 to 50 feet. The texture of both horizons on rock and alluvial parent material is quite clayey; so much so in fact, that the pallid horizons are mined for their kaolin content. Since the pallid horizons on the alluvial lateritic profiles are deeper, these are quarried by open-cut, but it has been found more economical to extract the kaolin from the shallower rock lateritic profiles by shaft mining.

Soil *colour* is less closely related to parent material than to the processes of soil formation; it is particularly influenced by whether humic acid or colloidal iron oxides have accumulated in the soil. Similar soil-forming processes can produce soils of similar colours even from contrasting parent materials. In some soils, especially those that are not fully differentiated, colour is inherited directly from the parent material. Carbonaceous shales tend to form black soils unless humus decomposition is very rapid. Rocks containing the less hydrated iron oxides tend to form reddish soils, unless the environment is very wet or the water table invades the profile. Conversely, rocks containing the more hydrated iron oxides tend to form yellowish soils, unless the environment is hot and dry and the water is driven out of the iron oxides. Inheritance accounts for some of the colour differences in the soils of the County of Cumberland (the Sydney Region) where red and yellow podzolic soils are found on shales and sandstones respectively.

Soil *texture* is very closely related to parent material, as was pointed out in Chapter 3. Parent materials containing minerals which weather readily tend to form clayey soils, while parent materials containing minerals that resist weathering tend to form sandy soils. Of course, a point is reached where the environment is so intense (for instance in the tropics) that almost all minerals are susceptible to chemical attack; then the parent material's influence on texture is rendered subordinate. Variation in texture with depth in soil profiles depends on the amount of colloid formed, and this is related to the ease of weathering of the parent material. There is a limit, which varies from environment to environment, beyond which so much colloid forms that illuviation is unimportant. This limit is exceeded in lateritic soils, krasnozems, black earths and prairie soils, but is not reached in the podzolised soils. At the other extreme, profiles lacking colloids are also uniformly textured[1].

Soil *structure* is closely linked to parent material because it is usually more closely related to the amount of colloid than to any other soil property. The best peds in soils are found on high colloid-forming parent materials such as basalt, while the low colloid-forming parent materials such as granite and sandstone usually have structure problems. In the New England area of New South Wales, fully pedal soils (90% to 100% in peds) on basalt may be found in the same paddock as apedal soils on granite.

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[1] The term *uniform texture* means that the same texture grading is used throughout the solum, e.g., the solum may be classsed as clay loam throughout. It does not imply that there is *no* texture change, e.g., the subsoil may contain 5% more clay than the topsoil and still fall into the same texture class.

Soil reaction and the presence of lime are more closely allied to climate than to parent material in so far as all parent materials can form acid or basic soils, with or without lime. But in a small area of uniform climate, relief and vegetation, *pH* differences can be related to parent material. The more clay-forming the parent material, the greater the likelihood of encountering lime in the profile and the higher the *pH*. Leaching is far more effective in light-textured soils.

The fertility of a soil is partly dependent on the underlying parent material, as mineral sources for some plant nutrients are more important than organic sources. Parent materials high in bases and containing the trace elements have a definite advantage over parent materials lacking these. A soil's reaction to cultivation is conditioned not only by its nutrient status but also by its physical constitution. Many soils from low colloid-forming parent materials (for instance, sandstone and granite in the New England area) have eroded so badly and degenerated in structure to such an extent that they have been abandoned.

RELIEF

Slope exerts a micro-climatic effect on the soil environment. On the upper slope, rainfall is most effective as a leaching agent because only fresh rainwater penetrates the soil. This water then moves laterally down the slope, carrying with it the salts dissolved from the upper slope soils; this water combines with the surface rainfall on lower slopes so that the resultant mixture is more highly charged with salts than water on the upper slope. This increase in concentration of the soil solution lowers the solubility of cations and anions in the lower slope soils, sometimes to such an extent that as soil water evaporates, salts are precipitated in the lower slope profiles. For this reason, lower slope soils generally have higher *pH* values than upper slope soils and are more likely to contain lime. Although upper slope soils dry out more rapidly than lower slope soils, and are therefore potentially more arid, profile morphology indicates a reverse effect in terms of leaching. Better drainage and more effective leaching are features of upper slope soils. Whenever there is a salting problem, it is always on the lower slopes, for it is here that the salts leached from the entire slope can accumulate. When one walks down a long slope of uniform parent material, one often encounters soils similar to those in a traverse across an area of decreasing rainfall. This is obvious when one considers the mid-slope soils on basalt from the coast to the inland across northern New South Wales. Krasnozems occupy whole slopes in areas receiving 60″ average annual rainfall. In areas receiving 40″ rainfall, krasnozems are found on the upper slopes, chocolate soils on the middle slopes, and prairie soils on the lower slopes. In areas receiving 30″ rainfall, krasnozems occupy the upper slopes only if they are steep; otherwise, chocolate soils extend down to the middle slope where prairie soils are found, with black earths on the lower slope. In areas receiving 20″ rainfall, black earths occupy the whole slope.

Because lower slope soils remain wetter for longer periods than upper slope soils, their iron oxides tend to become hydrated, so that yellow rather than red colours predominate. Upper slope soils which are usually freely drained, are more reddish in colour. Repeated water-logging of lower slope soils causes mottles to develop, so drab greys and yellows are usually interspersed. This, coupled with the greater likelihood of lime accumulation, makes lower slope soils much duller and paler in colour than their upper slope counterparts.

The steepness of the slope is no less important than the position of the soil on the slope. The steeper the slope, the greater the tendency for erosion and therefore the shallower the profile. The following graph, based on podzolic soils in the Mudgee area of New South Wales, indicates the relationship between steepness of slope and depth of the eluvial horizons (includes *A* and *E*

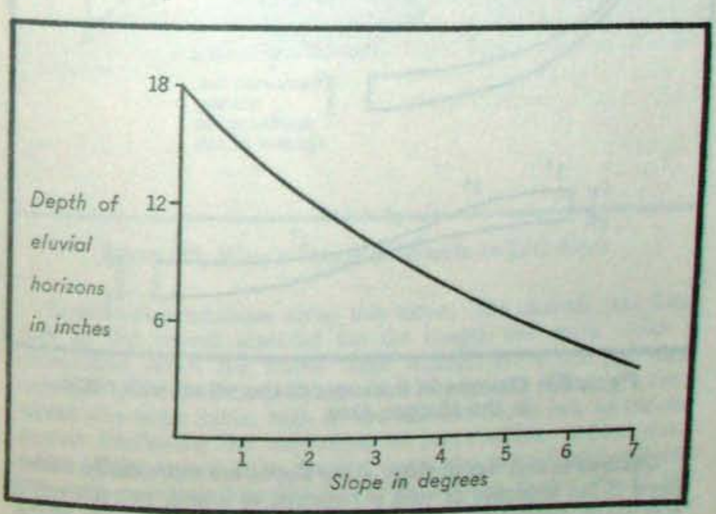

Figure 5.6: Relation between depth of eluvial horizons and slope steepness in the Mudgee area.

horizons). Rainfall averages 30″ per annum, and the parent material of the soils considered is shale and slate.

The relationship is not quite as simple as this graph implies, as slope position interacts with slope steepness to make lower slope soils deeper and upper slope soils shallower than would be the case if normal erosion did not operate. There is always a skinning of the topmost layers of upper slope soils and movement of these layers downslope, sometimes to be deposited on top of the soils of the lower slope. The combined effect of the two variables is apparent in the contrasting depths of soils along concave slopes (increasingly steeper slope with increasing height). The upper slope soils on the very steep slopes are often so shallow that they may be skeletal, while a considerable depth of soil may accumulate on the gentle lower slope. The following diagrams indicate the relationship between slope steepness and position in podzolic soils on slate in the Mudgee area. In these cases, slope steepness outweighs slope position:

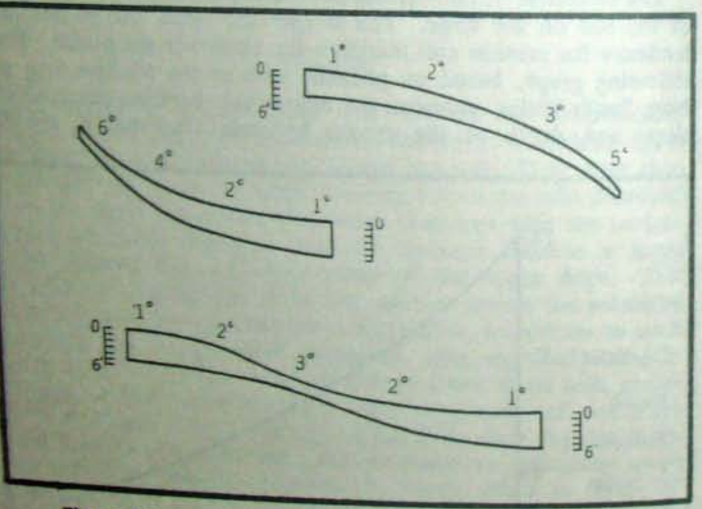

Figure 5.7: Changes in thickness of the solum with slope in the Mudgee area.

Changes in soil morphology along slopes are more easily understood if the sequence of soils is regarded as a unit, as each soil is genetically related to those on either side. In 1936, G. Milne, working on granitic soils on slopes in East Africa, introduced the term *catena* for the sequence of soils along a slope resulting from the processes of normal erosion and seepage. He likened the morphological features of the soils to the links of a chain, each link depending on those on either side of it, so that what happens to one link is transmitted to the whole chain, affecting immediate links first and to the greatest extent. Any change in properties of a soil on a slope will affect its neighbouring soils first and most noticeably. The change will be transmitted in diminishing intensity to all the soils along the slope until it has become evenly distributed or obliterated.

Milne illustrated his catena concept by describing the sequence of soils along a granite slope in the plateau region, south of Lake Victoria in Africa. His sequence is shown in Fig. 5.8:

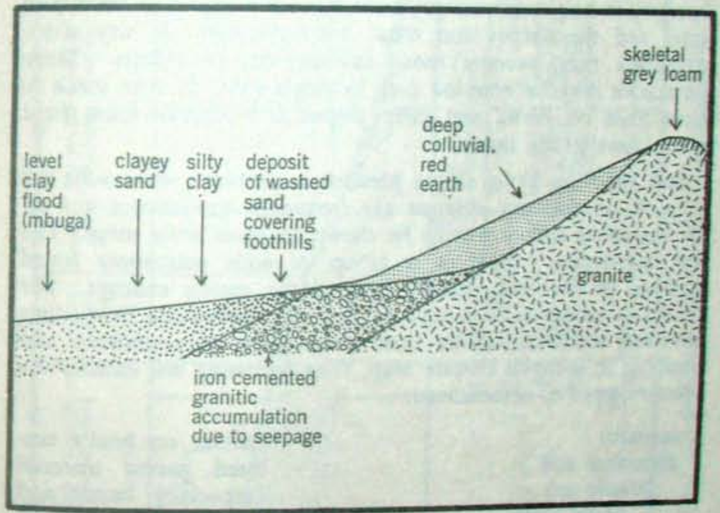

Figure 5.8: Milne's catena on granite in East Africa.

Erosion is continuous along this slope. The skeletal grey loam acts as the parent material for the deeper red earth which is colluviated down the slope. Iron leached from the red earth cements granitic and pedologic detritus at the break of slope where the water table, high in the wet season, is low in the dry season facilitating the deposition of sesquioxides. Storm water pares off the topsoil from the grey loam, the red colluvial soil and the iron accumulation; the resulting detritus moves across the lower slope and into the valley, to be deposited in order of particle size, the sand falling nearest the iron accumulation and the clay forming a wide flood, termed *mbuga*, across the valley floor.

The term catena was originally used for slopes of uniform parent rock (in the above case, granite, forming granitic colluvium), but its use has been extended to slopes with lithological variation (for instance, dipping sedimentary rocks of varying composition and texture), and to slopes which may have undergone more than one period of soil formation. The term catena is used also for slopes the lower parts of which are mantled with alluvium, possibly quite different in composition from the hillside soil parent material.

During the development of a catena, vegetation variation down slopes is concomitant with soil variation. One of the first premises which applies to even or concave slopes is that vegetation increases in height downslope in response to the greater amount of water and the deeper and more nutritious soil. In dry areas, vegetation may become more salt-tolerant downslope. These differences may be detected only in virgin soils, as man tends to leave trees on crests and steep slopes and remove them from lower, more gentle slopes.

Soil variation along slopes presents a problem when soils are mapped, as catenary changes are frequently gradational and all the variations cannot usually be shown on small scale maps. The *soil association*, which is a group of soils commonly found together in the field, usually involves the catena concept. The catena is mapped as a unit in the soil association map, thus obviating the need to draw lines where there are gradations, and resulting in a much simpler map. The following soil catenas are often mapped as associations:

krasnozem chocolate soil prairie soil black earth wiesenboden (meadow soil)	Catena on heavy textured parent material (especially basalt and basaltic alluvium)
alpine humus soil brown podzolic soil red and yellow podzolic soils grey-brown podzolic soils	Catena in the Monaro Region of New South Wales on variable parent material
red podzolic soil yellow podzolic soil gley podzolic soil	Catena in moist temperate regions, such as the County of Cumberland.

Every member of a catena need not be present, but the order is always the same; for instance, gley podzolics are not necessarily

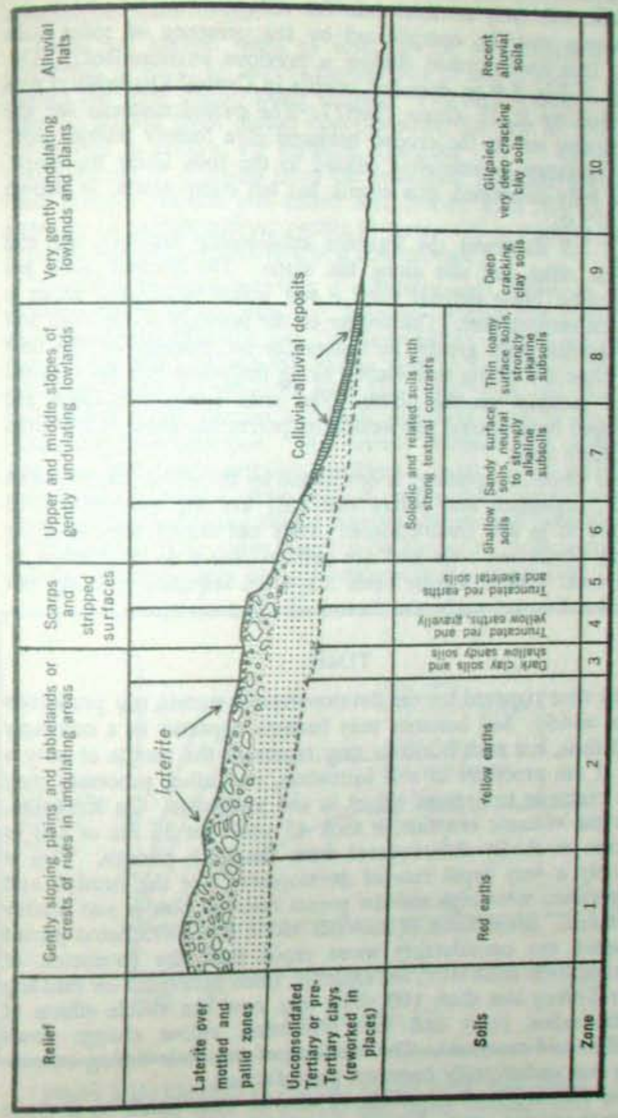

Figure 5.9: Soil catena on denuded laterite in Queensland (after R. H. Gunn, 1967).

the lowest member of a red-yellow podzolic catena, as yellow podzolic soils may continue into the valley.

Catenas may be complicated by the presence of relict soils (soils that have formed during a previous environment). The catena in Fig. 5.9 on denuded laterite in Central Queensland was described by R. H. Gunn (1967). The parent material for the present-day soil is the eroded horizons of a former lateritic soil. The vegetation is intimately related to the soils along the slope. When fully developed, this catena has ten components, as shown in Fig. 5.9.

Fig. 5.9 illustrates the intimate relationship between the soil profiles, relief and site along the slope. The surface soils are polygenetic, being derived from a soil which weathered under a different environment. The nature of the horizons of the old soil (the lateritic soil) greatly influences the morphology of the new soil. Note the strong colluviation along the scarp and the alluvial parent material on the plains. The soils along this slope are developed from several pre-weathered parent materials rather than from fresh rock.

Soils whose morphology is dominated by the above three factors (relief, organisms and parent material) are the *intrazonal* soils referred to in soil classifications. They cut across the *zonal*, or climate-dominated soils and are seldom shown in soil maps of continental extent. Larger scale maps are required to bring out these local variations in the factors of soil formation.

TIME

The time required for the development of certain soil properties varies widely. Soil horizons may become apparent in a relatively short time, but such horizons may represent the results of only a few of the processes of soil formation, and other processes may take centuries to become extant in soil properties. On Krakatau, after the volcanic eruption, it took 45 years for 35 cm of soil to become markedly differentiated from the fresh pumice. This is possibly a very rapid rate of development, for the rainfall and temperature were high and the parent material porous and readily weathered. Movements of material within a pre-weathered parent material are considerably more rapid than the formation of colloids from fresh rock; for example, shale excavated for building near Sydney less than 100 years ago, now has visible effects of podzolisation (clay and iron illuviation, colour change down profile, acid reaction). The breaking of the shale during excavation has undoubtedly hastened podzolisation.

In 20 or 30 years, stabilised coastal dune sands in a moist climate assume some of the features of a soil profile, such as the darkening of the surface few inches by humus, but illuviation of constituents takes longer.

There are several ways of studying the time taken for soil properties to form; all require the absolute dating of several points during the time period being studied. The rate of podzolisation may be studied from glacial deposits when the rate of retreat of the glacier is known; for instance, Chandler (1942) studied the podzolisation of moraine from the retreating Mendenhall Glacier near Juneau, Alaska, and found that it took from 500 to 1,000 years for a well-developed profile to form. Other workers believe it takes even longer than this in a cold climate.

Tree-ring analysis provides a relatively easy means of dating deposits from which soils are forming. Using this method of absolute dating, Dickson and Crocker (1954) studied soil formation of mudflows in the Mt. Shasta area of California. The five mudflows were 27, 60, 205, 566 and 1,200 years old, and they believed that the climate has been relatively constant since the oldest mudflow stopped. The parent material was weathered andesite and the vegetation coniferous forest. They found the colours of the top 2½ inches of soil on the three oldest mudflows were similar, indicating that a constant colour had been reached in 205 years. The lower layers took longer to reach a constant colour. Their findings are summarised in the graph (Fig. 5.10).

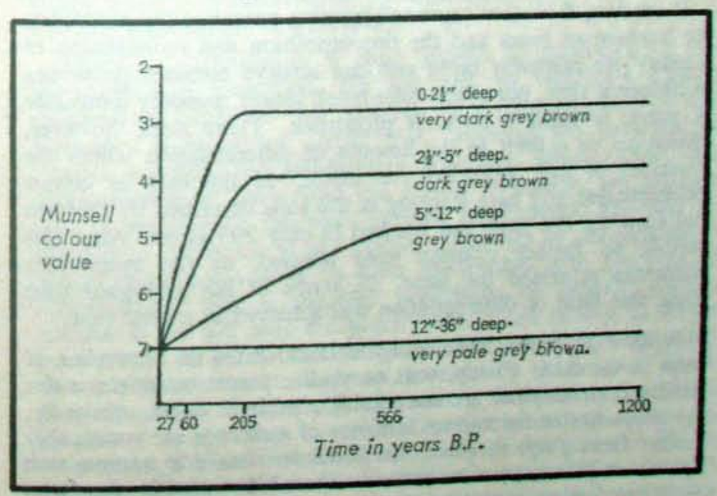

Figure 5.10: Changes in Munsell colour value with depth in soils of varying age (after Dickson and Crocker).

The colour change in these soils was due primarily to the decomposition of humus as treatment of the soils with hydrogen peroxide, which dissolves humus, changed their colour back to Munsell value 7. Reddish and yellowish colours were found only in the three oldest flows, indicating that not enough iron oxides accumulated to colour soils in 60 years.

Dickson and Crocker found that *pH* also reached a constant value in 566 years, as values for soils on the 1,200 year old mudflow were similar to values for soils on the 566 year old mudflow. However, other soil-forming processes were much slower. They found little evidence to suggest the formation of colloid in the 1,200 year period of soil formation, as the sand-sized minerals remained unaltered, except for basaltic glass. Slow rates of formation of soil colloids are generally accepted by pedologists. Barshad (1965) claims that only about 0.002 gm of clay form annually from 100 gm of parent material in soils in mid-western United States. However, this is perhaps not an inconsiderable amount in view of the fact that clay forms in all horizons simultaneously and may be translocated within the profile. It is rather surprising that Dickson and Crocker found no evidence of clay formation over such a long time. Many soils could have been weathering since the Tertiary, and this would have given them ample time to form a considerable amount of colloid.

It appears, then, that some soil-forming processes (in particular, the leaching of bases and the decomposition and mobilisation of humus) are relatively rapid and can achieve constant properties in historical time, but others take much longer, possibly thousands of years, to achieve constant properties. There does, however, appear to be a limit to the amount of differentiation which the processes of soil formation can effect. In the case of colour differentiation and base leaching in the soils described by Dickson and Crocker, the limit was reached in only 205 years. After this period, no further changes were effected, as the measurable properties remained the same. In terms of the geological time scale, this limit of differentiation was achieved in a brief time.

A limit to profile differentiation is exhibited in a number of ways in the field. Firstly, soils on similar parent materials, under similar environments, are remarkably similar in depth. Secondly, the properties of the various horizons of such soils are remarkably similar from place to place. It would be unsafe to assume that all the soils in question have weathered for exactly the same durations; rather, their ages are fortuitous. Instead of progressive soil differentiation with increasing time, it is more likely that, in fairly old soils, the current environment can deform the soil materials only to the extent attained, and is incapable of transforming them any further. Were this not so, one would be unable to select a single profile description to represent a given combination of parent material, relief, organisms and climate, but would have to describe every exposed soil in a given locality. Soil depth and degree of horizon differentiation would be directly proportional to the duration of weathering, and there would be little correlation between depth of soil and slope steepness: indeed, the deepest soils would be the oldest. Also, one would expect to find podzolised soils hundreds of feet deep if they have been weathering for thousands of years. The work discussed so far, especially Dickson and Crocker's, indicates that this is not so, and that changes in soil properties 'tail off' with increasing time after a certain minimum duration, during which the properties change progressively. This 'tailing off' may represent a hyperbolic function—perhaps changes are continually occurring but the pedologist's naked eye and instruments have not been sensitive enough to detect them.

This apparent lack of further differentiation of soils after they have reached their limit in depth and development of properties does not imply that the soil profile is passive; rather, the soil-forming processes are actively maintaining the horizons in their present morphological relationships. If they did not, the profile would soon be obliterated by the accumulation of matter and a resulting destruction of properties.

After the exposure of a fresh surface to processes of uninterrupted soil formation, the soil and its vegetation change progressively until the soil becomes fully differentiated and the vegetation is at climax. After that, providing the environment remains constant, the soil-forming processes, which include the decomposition of litter from the climax vegetation, are efficient only in maintaining the soil's present morphology, and are ineffective in deforming it any further. The rate of decomposition of leaf litter keeps such an even relationship with the rate of accumulation, that a constant percentage of organic carbon remains in the soil over the years. Such a soil profile, in which there are no further visible changes with increasing age, provided the environment remains constant, is termed a *mature soil*. Since the processes of soil formation are actively maintaining the soil in its present morphology, in spite of environmental tendencies to obliterate it, the soil is said to be in *dynamic equilibrium* with its environment. The soil profile constitutes an *open system* in which materials may enter at one end and similar materials leave at the other end without effecting any permanent change in the soil's properties.

How, it may be asked, can the geomorphic surface be lowered when the soil-forming processes are no longer bringing about any visible changes? The effects of soil erosion can be very spectacular, but not all removal of material is so spectacular, and few observers realise how important can be the effects of a gradual skinning of the upper part of the soil profile. The vegetation may not be disturbed, but even at the bottom of grassy slopes there will probably be found, after heavy rain, a thin layer of newly deposited soil which soon becomes part of the soil beneath as the grass grows through it—a thin top-dressing. This less spectacular form of soil erosion and deposition can operate continually without permanently changing the depths of either the eroding or accumulating soil. In this way, the landscape and the soil with it may continually be lowered, the eroded material being either carried to rivers and then oceans, or deposited on lower slopes or in the valley. Thus dynamic equilibrium in soils forms an integral part of the lowering or retreat of equilibrium surfaces in geomorphology.

C. C. Nikiforoff, an American pedologist, first drew attention to the concepts of the mature soil and dynamic equilibrium in soils in a series of papers published during the 1940s. He considered that there are two types of mature soil, *cumulative* and *non-cumulative*, which are gradually being buried and eroded respectively. Neither of these soils changes over time. They can both occur on the one slope, as shown in Fig. 5.11; or they can evolve independently, as when the material eroded from a non-cumulative soil completely leaves the area by streams; or a cumulative soil can receive accessions of materials from the erosion of a non-soil material such as talus.

Let us consider hypothetically the evolution of cumulative and non-cumulative soils on the one slope. In the non-cumulative soil, the steady skinning of the upper *A* horizon is offset by a proportional leaching of the upper *B* horizon; that is, the upper *B* horizon now becomes lower *A* horizon. Similarly, the upper *C* horizon becomes lower *B* horizon by the deeper illuviation of constituents, while a proportional amount of fresh rock weathers to form lower *C* horizon. Thus there is a progressive lowering of the soil horizons as shown in Fig. 5.11. In the cumulative soil, there is a slow, steady burial of the *A* horizon by material from the upper slope non-cumulative soil. The lower part of the *A* horizon is then too deep to be leached, so constituents are illuviated into it from above. Thus lower *A* horizon becomes upper *B* horizon and lower *B* horizon, now too deep to receive illuviated constituents, degenerates into material resembling *C* horizon material—the shiny clay skins around the peds, kept lustrous by the continual movement of soil colloids, are too deep to be affected by soil-forming processes, and so lose their character. Roots and microbial activity may no longer function at this depth. All of this causes soil structure degeneration.

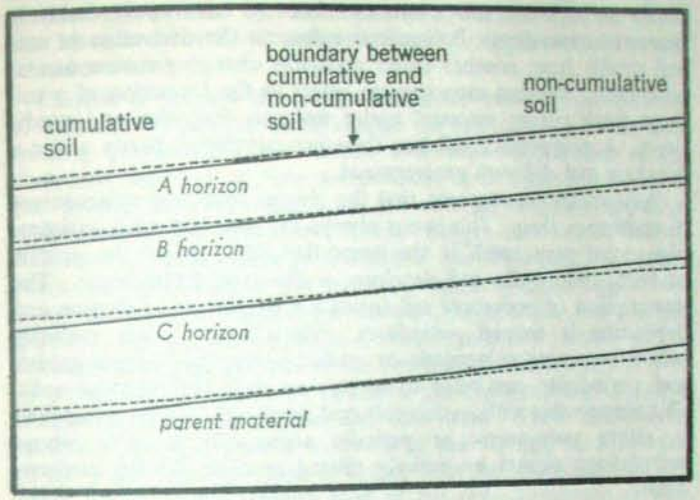

Figure 5.11: The formation of cumulative and non-cumulative soils.

The diagram in Fig. 5.11 represents a model only and one would be unlikely to encounter such a simple case in reality. Moreover, the diagram tends to imply down-wearing rather than back-wearing in its present form of gross exaggeration of scale and slope. This implication is not intended; rather, the cumulative and non-cumulative soils form parts of constant-slope features such as pediments and their associated veneers. Some of the podzolic soils in the Mudgee area do appear to have evolved in the manner indicated in Fig. 5.11, with bedrock exposed immediately beneath the *C* horizon of the shallow upper slope soil, while the deeper lower slope soil is underlain by deeply weathered material, its resemblance to the *B* horizon diminishing with increasing distance from it.

In Fig. 5.11, one presupposes that the environment remains constant and that erosion and deposition are sufficiently slow for soil formation to keep pace. Nikiforoff was fully aware that this was not always so and that the environment may change either gradually or catastrophically.

A gradual change in the soil's environment (such as a progressive increase in rainfall, without increased storminess, or a slow drop in temperature causing a slow increase in the effective precipitation by reducing evaporation) may cause the old soil to be gradually transformed into a different type of soil. Nikiforoff refers to this as soil evolution, but the term *polygenesis* is preferred nowadays. Polygenesis refers to the derivation of one soil profile from another under gradually changing environmental conditions, whereas monogenesis refers to the formation of a soil from fresh parent material under uniform environmental conditions. A polygenetic soil has therefore developed partly under a previous and different environment.

Polygenesis presupposes that the change from one environment to another is slow. This is not always the case and some catastrophic event may result in the truncation or burial of the profile, or both, before the soil develops in the new environment. The interruption of periods of soil formation by periods of erosion and deposition is termed *periodicity*. Not all soils reach maturity before becoming polygenetic or eroded or buried, but polygenesis and periodicity can refer to mature as well as immature soils. The terms *relict soils, palaeosols* and *fossil soils* are terms applied to either polygenetic or periodic soils, that is, soils whose morphology cannot be entirely related to their existing environment.

POLYGENESIS

The term polygenesis is more commonly applied to the derivation of one soil profile from another, under slowly changing environmental conditions, without erosion of the older profile. Some pedologists use the term for soils weathered from the truncated horizons of old profiles, but these will be included under 'Periodicity' in this book, because during the truncation of the old horizons soil formation must have been interrupted. Here, the term polygenesis will be used only when soil formation continually dominates erosion-deposition, under a changing environment. During the change in environment, some of the old processes are slowly replaced by new ones, producing different morphological characteristics in the new soil. But the progressive change in the soil's environment is so slow that the processes of soil formation can keep pace, though they may themselves change. Polygenesis bears some analogy to antecedence in streams, in which uplift (the changing environment) is sufficiently slow for down-cutting to keep pace. The new stream channel, like the polygenetic soil, will undoubtedly differ from the channel prior to antecedence.

Polygenesis clearly illustrates Stephens' integral equation relating the factors of soil formation (see the introduction to this chapter).

The soils we see today may have undergone a complex history of changes in one or several of the factors of soil formation over time. Certainly where the dependent factors are involved, the change would have been in several factors; for example, in climate and vegetation or relief and parent material. Polygenesis may therefore be initiated by variation, over time, in any of the other four factors of soil formation—climate, organisms, relief and parent material.

Climate is perhaps the most widespread cause of polygenesis, as slow, steady climatic shifts change the rate of most of the soil-forming processes. For example, a decrease in rainfall causes decreased leaching, so lime accumulates at a shallower depth in the profile, the *pH* rises, and the rate of colloid formation decreases—to list just a few of the changes that would follow. Evidence for polygenesis is difficult, if not impossible, to detect when the shift is from less to more intense weathering, as the latter tends to obliterate the former. But when the reverse happens, that is, when there is a shift from more intense to less intense weathering, features of the previous soil are more likely to be preserved. Consider, for example, the changes transforming a chernozem into a prairie soil, and vice versa. Under diminishing rainfall, the new climate is incapable of maintaining the full depth of the prairie profile, which is deeper than the chernozem, so the lower layers do not become part of the new chernozem profile, in which lime accumulates within the solum in the *B* horizon. Under increasing rainfall, the chernozem profile is deepened and lime is leached from the solum to the *C* horizon to form a prairie soil. In this case the properties of the old soil (the chernozem) are obliterated, but in the previous case some of the prairie soil remains beneath the chernozem as evidence of the earlier, wetter climate.

Increased aridity is often accompanied by an increase in salt in soil profiles because the new rainfall is less effective in removing the salt. In southern Australia, salt from the sea is continually being carried across the continent by the cyclonic systems. In arid periods this salt accumulates in soil profiles. In wetter periods some, but perhaps not all of the salt may be removed, and the remaining sodium and magnesium will be adsorbed onto the soil colloids as evidence of the previous drier climate. Since mineral matter is actually added to the soil profiles, the accumulation of salt (called *cyclic salt*) may be attributed partly to climatic variation and partly to the addition of new *parent material* (the salt).

Climatic change is accompanied by *vegetation* changes, and the combination of the two factors produces changes in the soil profile.

However, man has interfered with natural vegetation for such a long time that changes in the organism factor may be quite independent of climate. Man's massive clearing of trees and their replacement by grass for grazing has caused marked morphological changes in the associated soils. Not infrequently this drastic change has been catastrophic, causing severe erosion and therefore resulting in periodicity rather than polygenesis.

Polygenesis is quite common when the level of the *water table* changes either naturally or artificially. Invasion of a well-drained soil by the water table results in prominent subsoil mottling, which, after the water has returned to its previous level, takes either a long time to disappear or does not disappear at all.

Although polygenesis is perhaps more difficult to detect in the field than periodicity, many Australian soils that were previously thought to be monogenetic are now considered to be polygenetic. The light textured red and brown soils associated with the Cobar pediplain are classic polygenetic soils, as they are deeper and their *pH* is lower than neighbouring monogenetic soils. These soils have an intense reddish or brown colour indicative of the mobilisation of iron oxides in a previously wetter climate. The solodic soils of New South Wales, which resemble podzolic soils in their morphology, but contain more exchangeable sodium and magnesium than monogenetic podzolic soils, are also polygenetic. These soils, which are found extensively on the drier side of podzolic soils, are slightly deeper than neighbouring podzolics. These and other polygenetic soils are described in Chapter 12 which deals with relict soils in Australia.

PERIODICITY

The alternation of periods of soil formation with periods of erosion-deposition has long been recognised, especially in Australia, where knowledge of soil periodicity is well developed. The first evidence of periodicity came from buried soils which are common everywhere in Australia. Deep cuttings in roads, railways and quarries frequently expose successions of buried soils some of which are quite different morphologically from the present-day soil on top, and therefore reflect a previous and different environment.

The occurrence of buried soils in vertical succession is termed *soil stratigraphy* and the techniques used in this field are analogous to those used in rock stratigraphy—techniques guided especially by the principle that 'the present is the key to the past'. Studies in soil stratigraphy appeared quite early in Australian pedological literature. Woolnough (1918 and 1927) suggested that duricrust (indurated material, divided according to its composition into silcrete, laterite and calcrete, and formed during intensive weathering in the Tertiary) be used as a marker for arranging the soils above and below it in stratigraphical sequence, as in geological stratigraphy. But duricrust poses greater problems than rock strata, as it represents the oldest soil formation so far found in Australia, and most of the younger soils form at lower levels than the duricrust. Duricrust is very common in Australia. In the east it is expressed as laterite, the iron-rich component. This hard, red and yellow material should be looked for in railway and road cuttings and in quarries. In the dry interior, it is more common as silcrete (the silica-rich component) or calcrete (the calcium carbonate-rich component). Duricrust is often the first indicator of soil periodicity that one finds in an area. More will be said about it in Chapter 12.

The word *periodicity* was used initially by J. Macdonald Holmes to describe a soil succession in the Tamworth area in 1937 and 1940. He stressed the importance of 'deposit morphometry' in the study of soils and he particularly related periodicity to Milne's concept of the catena. Periodicity then received very little attention in Australia until the start of the landscape studies in the Riverine Plains in the early 1950s. Butler's (1959) paper on Periodic Phenomena in Landscapes renewed interest in the subject in Australia and was followed by a series of papers, chiefly by C.S.I.R.O. pedologists, on soil stratigraphy throughout Australia. Butler (1959) and his co-workers used *K-cycles* in their studies. Butler defines *K*-cycles as 'the alternation of phases of progressive soil development and minimal surface succession with phases of progressive surface succession and minimal soil development'. His *K*-cycle starts with an unstable phase in which the old soil is eroded and/or buried and soil development is minimal. The unstable phase is followed by a stable phase when the new parent material undergoes soil formation and erosion-deposition is minimal. A *K*-cycle, then, constitutes one unstable and one stable phase and ends with the onset of the next unstable phase. Butler uses the term *groundsurface* for the material formed during one *K*-cycle, that is, a soil and its parent material. *K*-cycles are named backwards from the present as follows:

K_{0u} present unstable phase
K_{1s} present soils
K_{1u} the erosion-deposition of the parent material of the present soils
K_{2s} soils of the environment immediately before the present, etc.

K-cycle nomenclature has lately been abandoned, partly because of the implied correlation between *K*-cycles of different areas. *K*-cycle chronology is relative and applies only to the area in which it was formulated. Correlating *K*-cycles without absolute dating is extremley dangerous. Further comments on *K*-cycles will be made in Chapter 12 when the relict soils of the Riverine Plain are considered.

Soil stratigraphy is analogous to rock stratigraphy in the unconformities and disconformities that are found. Irregular truncation, such as that arising from pediplanation, exposes an unconformable surface for soil formation during the period following truncation. The horizons of the truncated soil profile may serve as the parent material for the soil formed during the ensuing stable phase; for example, the horizons of a lateritic soil may be the parent material for red and yellow podzolic soils as shown in Fig. 5.12. The truncated horizons may be buried by either colluvium or alluvium or both, and the new soil may form on these fresh parent materials (see Fig. 5.13).

Truncated lateritic profiles are widespread in Australia and are the parent materials for many Great Soil Groups (see Chapters 6 and 7). Since lateritic profiles are deeply weathered, it takes a great deal of erosion to destroy them completely, whereas shallower profiles such as podzolic soils or black earths are obliterated in a much shorter time or with much less intense weathering.

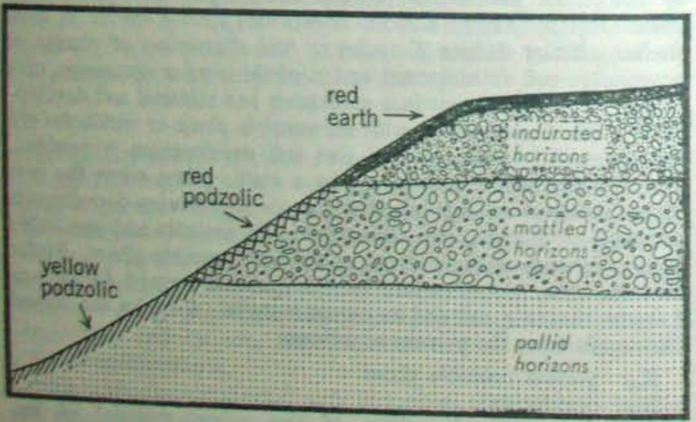

Figure 5.12: A truncated lateritic profile forming the parent material for later soils. The reddish indurated horizons form a red earth (krasnozem), the red and white mottled horizons form a red podzolic and the whitish pallid horizons a yellow podzolic soil.

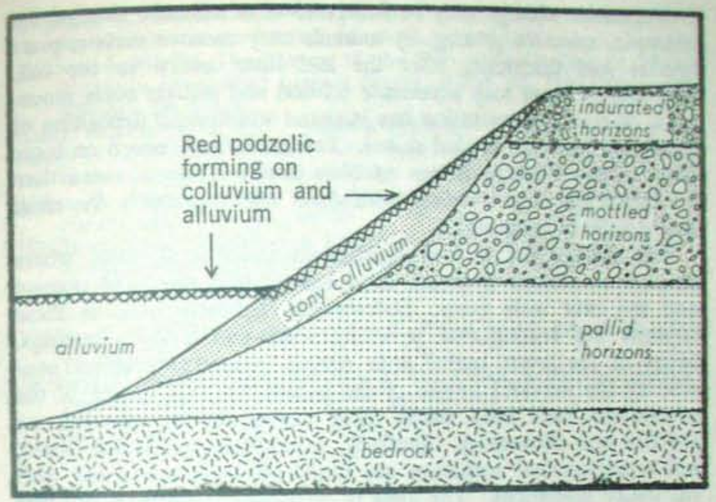

Figure 5.13: A truncated lateritic profile buried beneath colluvium and alluvium which are developing a red podzolic profile.

Soil stratigraphy, like rock stratigraphy, indicates a great deal about the environment in which a soil formed, for pedologists know the environments under which the various Great Soil Groups form today. If an area contains a relict soil which is actively forming today but only under hotter, wetter conditions than those where it is found, one must infer that the area had a previous hotter, wetter climate. Thus relict soils assist the palaeoclimatologist (the student of previous climates).

Climatic change is the most widespread cause of soil periodicity. Climatic change is often associated with instability and increased storminess, which accelerates erosion, causing soil profiles to be truncated. The eroded material may be deposited elsewhere to bury other soils. This constitutes an unstable phase, so that, after the new climate has settled in, a mixed surface of truncated soils and fresh parent material presents itself for soil formation.

A decrease in temperature has a twofold effect on soils. Firstly, effective precipitation increases, so there is more leaching; secondly, there is greater likelihood of frost with its associated upheavals of the soil and accelerated mass movement, especially solifluction. The onset of aridity induces instability through devegetation and an increase in aeolian activity.

Vegetation change may be independent of climatic change; for example, selective grazing by animals may remove certain plant species and drastically alter the leaf litter added to the soil. Removal of grass may accelerate erosion and initiate mass movement of soil. Deforestation has initiated widespread deposition of colluvium from the eroded slopes. This colluvium, which on some parts of the Western Slopes of New South Wales is more than 15 feet deep, is now being podzolised and represents the most recent soils in some areas.

Soil stratigraphy is well established in New Zealand where successive volcanic eruptions have spread thin layers of pumice and ash over wide areas. Successions of buried soils in these deposits are encountered in many road and railway cuttings. Many of the lower buried soils remain remarkably intact considering the pervious nature of the parent material. Some of the successions have been dated using carbon–14 to establish an absolute soil chronology. Similarly, in midwestern United States, soil stratigraphical sequences are found on loess associated with the major glaciations. The loess is calcareous, so one of the first soil-forming processes is the leaching of the lime and the lowering of the *pH*. In Australia, buried soils are found on a wide variety of substrates—aeolian material, colluvium, alluvium, dune sands and coastal depositional features. The student should begin at once to look for them in cuttings, creeks and quarries, enquiring whether they differ from the present-day soils of the area, and whether differences relate to their parent material or to climate.

A deep soil succession may represent many combinations of the four factors of soil formation (climate, organisms, parent material and relief) over time before the present-day period of soil formation commenced. It is the pedologist's job to determine what these factors were at the various points in time when soil formation was greater than erosion-deposition, that is, when the various buried profiles in the succession formed.

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Chapter Six

Soil Classification

CONTENTS

> 'Classifications are contrivances made by men to best suit their purposes. They are not themselves truths that can be discovered. Therefore there is no true classification; . . . the best classification is that which best serves the purpose or purposes for which it was made.
>
> As knowledge expands, new facts or closer approximations to truth make new classifications imperative. Classification is the mirror in which the present condition of the science is reflected; a series of classifications reflect the phases of its development.'
>
> Guy Smith, in *Soil Classification, a Comprehensive System, Seventh Approximation* (U.S. Department of Agriculture, 1960).

THE REMAINING CHAPTERS IN THIS BOOK DEAL WITH GROUPS OF soils dominated by common processes of soil formation. The soils described in each of the ensuing chapters have certain common morphologic features resulting from these common processes. In order to place the world's soil spectrum into manageable categories, some form of classification is essential. The classification adopted in this book is based on those soil properties which reflect the more common processes of soil formation. Before examining the classification used here, let us first examine some of the principles of soil classification.

APPROACHES TO SOIL CLASSIFICATION

There are three main approaches to the classification of soils, and each is used for a particular purpose. First, there is the empirical or *taxonomic* approach which is based on individual soil properties, such as shear strength (the cohesiveness of the soil, measured by the force needed to rupture the soil) or texture. Empirical classifications are most useful to engineers, who are less interested in the soil profile and soil genesis than in the soil's reaction to stresses and water percolation. Of more use to the pedologist is the second, the *morphologic* approach, which takes into consideration total profile features, with emphasis on field descriptions of soil profiles. Since soil morphology reflects the soil-forming processes, the classification used in this book is a morphologic one. A third approach to soil classification considers the factors which influence soil formation. Since the origin of soils is of paramount importance in this approach, it is referred to as the *genetic* approach. In this approach, soil properties are related to the five factors of soil formation (climate, relief, parent material, organisms and time). This is one of the oldest approaches to soil classification—the early Russian pedologists stressed the close correlation between soil types and climatic regions, while later workers, notably Sibirtzev, noted the influence of other factors on soil formation and introduced these into their classifications. The grouping of soils according to the soil-forming factors into zonal (climate), intrazonal (relief, parent material and vegetation) and azonal (time) dates back to last century and is still used in many modern soil classifications.

It is difficult to categorise modern soil classifications as empirical, morphologic or genetic, as frequently all three approaches are combined. Some classifications employ the factors of soil formation (genetic) as major divisions, and subdivide according to profile features (morphologic) and then according to individual soil properties (empirical).

The approach to classifying soils is similar in many respects to the approaches used to classify other natural objects, such as plants and animals. However, the classificatory terms are not always used in a common sense by all pedologists, perhaps because soil classification is younger than either botanical or zoological classifications and is therefore constantly changing. The table shown in Fig. 6.1 illustrates the parallel between botanical and soil classification.

Botanical Classification		**Soil Classification**	
Phylum	Spermatophyta	Order	Zonal
Order	Angiospermae	Suborder	Podzolised
Class	Dicotyledoneae	Great Soil Group	Red podzolic
Family	Myrtaceae	Association (Aust.) Family (Amer.)	Cumberland
Genus	Eucalyptus	Series	Cumberland
Species	obliqua	Class	clay loam
Common Name: messmate		Common Name: Cumberland clay loam	

Figure 6.1: A comparison of botanical and soil classifications.

In both systems of classification in Fig. 6.1, there are six levels and the common name is related to the last two levels. However, in the soil classification, the common name refers to the soil of a very localised area while the Great Soil Group is the more ubiquitous form. In botanical classification, families are perhaps almost as ubiquitous as Great Soil Groups are in soil classification. The terms in soil classification, from Family to Class refer to local soils, and such levels of classification would not appear on a world map of soils, but only on quite large-scale maps. Hence, this book is not concerned with levels of classification at or below the Family level.

Since soil classification is less firmly established than either botanical or zoological classification, the terms as outlined in Fig. 6.1 are not used in this sense by all pedologists. However, the order of the terms in the classification hierarchy is the same. Some pedologists even use different terms for the various levels of classification and introduce additional terms. This will be more obvious after some well-known soil classifications have been examined. First let us look more closely at the levels of classification of soils as shown in Fig. 6.1.

Soil *orders* usually relate to the five factors of soil formation. The soil order is the highest category in soil classification and usually refers to the division into zonal, intrazonal and azonal soils. Zonal soils have well-developed profiles and reflect continental variations in climate and associated vegetation. They develop on well-drained uplands on parent materials not extreme in texture or chemical composition. Intrazonal soils, which also have well-developed profiles, reflect the dominating influence of one of the other factors of soil formation, especially relief, parent material, or local peculiarities of vegetation. Intrazonal soils may form catenas with zonal soils and so may be of continental extent, but more often they reflect some local condition which would not appear on a map of continental extent. Azonal soils lack profile differentiation, either because they have not had sufficient time to develop or because they occur on slopes so steep that erosion balances soil formation to the extent that no horizons can form. In both cases, soils lacking profile differentiation will result. Although azonal soils on steep slopes may be quite mature in terms of their landscape, both types of azonal soil appear to be immature morphologically.

Some pedologists use soil orders to divide soils into pedalfers (iron-accumulating, from Latin *ferrum*, meaning iron) and pedocals (lime-accumulating, from *cal*, short for calcium). This division cuts across the zonal, intrazonal, azonal classification, but it still broadly reflects climate, both continental and local, and so is also related to a factor of soil formation.

Soil *suborders* are generally associated with the most obvious profile features, such as domination by organic matter, accumulation of sesquioxides, marked illuviation of constituents or the accumulation of salts. Zonal suborders are of continental extent, but intrazonal and azonal suborders may refer to local variations and so appear only on larger scale maps.

Great Soil Groups are based on total profile features. Since soil properties form a continuum, Great Soil Groups frequently grade into each other; for example, if two Great Soil Groups are distinguished mainly by colour, as in the red and yellow podzolic groups, it is likely that between these two main Great Soil Groups one will encounter soils with both red and yellow (or orange) colours. Great Soil Groups, then, represent nodes along the long continuum of soil properties. This is the sense in which this category of classification is used in this book, but there are intergrading soils between the main Great Soil Groups. Examples of Great Soil Groups are red podzolics, chernozems, red-brown earths and grey soils of heavy texture. These names are analogous to the names of common rocks in geology, such as granite, sandstone, basalt and slate. These names have certain meanings in common among geologists, but the precise definition of each term may vary; for example, the limiting percentage of minerals in a rock defined as granite varies among geologists, yet all geologists consider granite as an acid, igneous, plutonic rock containing quartz, feldspar and biotite and sometimes hornblende. Similarly, to the pedologist, red podzolic means an acid, markedly differentiated soil in which iron and clay have been illuviated and an *E* horizon has formed. The precise amount of illuviation considered necessary before a soil can be termed a red podzolic varies among pedologists. Great Soil Group names are the common pedological language, and though some pedologists are attempting to eliminate them from their soil classifications, they still represent the most concise way of describing soils in terms which most pedologists understand.

SOME SYSTEMS OF SOIL CLASSIFICATION

Some well-known systems of soil classification will now be considered. Three of these are by Australian pedologists (Stephens, 1962; Northcote, 1965; Stace, Hubble and others, 1968), one by an American pedologist (Guy Smith, 1960), and another is the FAO/UNESCO International System (1968). None of these five systems use the soil classification categories in the same sense. Stephens uses his system in *A Manual of*

Australian Soils which describes the morphology, genesis, occurrence and utilisation of the main Great Soil Groups recognised in Australia. His emphasis, therefore, is on classification down to the Great Soil Group level although he does consider the lower levels of classification. His book has been replaced by *A Handbook of Australian Soils* (1968) by Stace, Hubble and others, in which they introduce a different system of soil classification, combining and deleting some of the Great Soil Groups used by Stephens, and introducing new ones. Their classification is also a means of arranging material in their book, rather than an end in itself. Northcote's and Guy Smith's systems, on the other hand, are more classification for the purpose of describing and naming, and for mapping and survey purposes, though the authors hope their systems will achieve wider recognition. Northcote's system forms the basis for *The Atlas of Australian Soils* which, in 12 map-sheets, each at a scale of 1:2,000,000, provides a soil map for the whole of Australia. This series, completed in 1968, is probably in school libraries; it is essential to understand the system of classification on which it is based if the mapping units are to be meaningful. Northcote's system, which will be revised (the revision beginning in 1969) is intended only for use with Australian soils. Guy Smith's system, which encompasses the work of the U.S. Department of Agriculture, is meant to include all the world's soils, and is also being revised. The 1960 version of Guy Smith's system is described as the 'Seventh Approximation'.

The FAO/UNESCO system of soil classification uses combinations of many recognised soil classifications, including Stephens' (1962) and Guy Smith's (1960). The classification is designed for the proposed 'FAO/UNESCO Soil Map of the World', publication of which is scheduled to begin in 1969. This project is intended to cover every continent, at a scale of 1:5,000,000. When this series is complete, school libraries will undoubtedly obtain a copy. Draft copies of all the maps were presented for criticism at the Ninth International Soils Congress at Adelaide in 1968.

The names used for the soils (as well as the classification categories) vary in these classifications. Stephens, and Stace, Hubble and others, use well-established Great Soil Group names in their books. The other levels of classification are similar to some of the very early classifications, especially Stephens' system, which is similar to the classifications used by C. F. Marbut. Soil suborders are the only other level of classification used by Stace, Hubble and others. Northcote and Guy Smith, in their classifications, introduce completely new systems of nomenclature for soils. Northcote uses letters and numbers, rather than words, to denote

his soils, and Guy Smith uses a system of names, based mainly on Latin derivatives; none of his soil names are familiar pedological terms, though some of the names are now used in pedological literature outside the context of the Seventh Approximation. The FAO/UNESCO system uses a combined terminology—some traditional Great Soil Group names are used, together with new terms, some of which are taken from more recent classifications such as the Seventh Approximation.

Since Stace, Hubble and others (1968) and FAO/UNESCO (1968) use only two levels of classification in their systems, namely suborders and Great Soil Groups, the trend in soil classification appears to be away from a complex hierarchy. These two systems, however, are designed for the purpose of describing and mapping soils at Great Soil Group level at very small scales, and so are not concerned with finer levels of classification which are needed to describe and map soils at much larger scales. The systems of Stephens, Northcote and Guy Smith are aimed at describing and mapping soils at all scales, so these authors cannot really avoid complex classification hierarchies. The classification categories used by these three authors are shown in Fig. 6.2.

Stephens, 1962	Northcote, 1965	Guy Smith, 1960 Seventh Approximation
Solum class	Primary profile form (Division)	Soil order
Soil order	Subdivision	Soil suborder
Soil suborder	Section	Great group
Great Soil Group	Subsection	Subgroup
Family	Class	Family
Series	Subclass	Series
Type	Principal profile form	Type

Figure 6.2: Categories used in soil classification by Stephens (1962), Northcote (1965) and Guy Smith (1960). The highest category is listed at the top of each column.

Stephens' System (1962)

Although in his *Manual of Australian Soils* Stephens uses classification levels only down to Great Soil Groups, he explains the other terms in his classification hierarchy. His first division is into *solum classes*, depending on whether horizons are present or not. There are two solum classes:

I Solum undifferentiated, which represents azonal soils;

II Solum differentiated, which includes zonal and intrazonal soils.

I Solum class—undifferentiated

1. Alluvial Soils
2. Skeletal soils
3. Calcareous coastal sands

II Solum class——differentiated

PEDALFERS

a. Solum dominated by acid peat or peaty eluvial horizon

4. Moor peats
5. Alpine humus soils
6. Moor podzol peats
7. Acid swamp soils

b. Solum acid, and with organic, sesquioxide and sometimes clay illuvial horizons

8. Podzols
9. Ground-water podzols

c. Solum acid, and with clay and sesquioxide illuvial horizons

10. Lateritic podzolic soils
11. Grey-brown podzolic soils
12. Brown podzolic soils
13. Red podzolic soils
14. Yellow podzolic soils
15. Meadow podzolic soils
16. Non-calcic brown soils

d. Solum acid to neutral and lacking pronounced eluviation of clay

17. Yellow earths
18. Krasnozems
19. Lateritic krasnozems
20. Lateritic red earths
21. Terra rossa
22. Prairie soils

PEDOCALS

e. Solum dark coloured and slightly acid to neutral in eluvial horizons, calcareous illuvial horizons

23. Black earths
24. Wiesenboden
25. Brown forest soils
26. Rendzinas
27. Ground-water rendzinas
28. Fen soils

f. Solum saline or showing post-saline structure in the illuvial horizon

29. Solonchaks
30. Solonetz
31. Solodised solonetz
32. Soloths
33. Solonized brown soils

g. Solum with slightly acid to neutral eluvial horizons and calcareous illuvial horizons

34. Red-brown earths
35. Brown earths
36. Brown soils of light texture
37. Arid red earths
38. Grey calcareous soils

h. Solum with neutral to alkaline weakly developed eluvial horizons and calcareous and/or gypseous illuvial horizons

39. Grey soils of heavy texture
40. Brown soils of heavy texture

i. Solum with deflated, slightly acid to alkaline, eluvial horizons, and calcareous and/or gypseous illuvial horizons

41. Desert loams
42. Grey-brown and red calcareous desert soils
43. Red and brown hardpan soils
44. Desert sand plain soils
45. Calcareous lateritic soils
46. Stony desert tableland soils
47. Desert sandhills

Soil orders are Pedalfers and Pedocals.
Soil suborders are a to i.
Great Soil Groups are 1 to 47.

Figure 6.3: Stephens' 1962 system of soil classification, showing solum classes, soil orders, soil suborders, and Great Soil Groups.

Stephens' *soil orders* depend on whether lime and/or gypsum are found in the solum, regardless of whether they occur in the *C* horizon or parent material; hence there are two soil orders—pedocals (containing lime) and pedalfers (not containing lime). *Soil suborders* depend on the most obvious profile features such as the positions of horizons of organic matter, clay, sesquioxides, lime and gypsum. Stephens recognises nine suborders (*a* to *i* in Fig. 6.3).

Stephens' *Great Soil Groups* depend on total profile features, especially changes in colour and texture, and features related to poor drainage, lime, iron, organic matter, salt accumulation and relict properties. Stephens regards Great Soil Groups as nodes along a continuum, rather like mounds of sand in a sand tray: the mounds represent modal soils with full development of the properties concerned, while the flanks and basal sand represent the intergrading soils with less than full development of these properties as they merge into different properties related to the next Great Soil Group. Stephens recognises 47 Great Soil Groups in his 1962 classification (1 to 47 in Fig. 6.3), a slight increase over the Great Soil Groups listed in the earlier editions (1953 and 1956) of his Manual. Stephens' classification applies only to Australian soils and many more Great Soil Groups are recognised in world soil classifications.

Stephens uses the remaining terms in his classification hierarchy in a manner similar to that of many Australian, and some overseas, soil surveyors. *Soil families*, a more recent level of soil classification, are associated with differences in depth of the solum, and differences in properties of the *B* horizons, especially structure and mottling. For example, various red podzolic soils could be divided into families by solum depth, and by the presence of pedal or massive *B* horizons. *Soil series* relate to differences in parent material. These differences do not have sufficient influence on soil morphology to form a different Great Soil Group, but cause variations in colour and texture within a single Great Soil Group. For example, in the Sydney region, red podzolics are found on both shale and sandstone, giving rise to series variations. Both of the red podzolics have acid profiles and clay and iron illuviation, but the sandstone podzolics tend to be more coarsely textured and they are therefore more poorly structured than the shale podzolics.

Soil types depend on differences in the texture of the *A* and *E* horizons, and, with accompanying minor differences, in the *B* and *C* horizons. The soils mapped as different types usually have a common parent material.

From the family to type levels of classification, soils are given local names and may be shown on quite large scale maps. Such

1. No profile differentiation	1. Solonchaks
	2. Alluvial soils
	3. Lithosols
	4. Calcareous sands
	5. Siliceous sands
	6. Earthy sands
2. Minimal profile development	7. Grey, brown and red calcareous soils
	8. Desert loams
	9. Red and brown hardpan soils
	10. Grey, brown and red clays
3. Dark soils	11. Black earths
	12. Rendzinas
	13. Chernozems
	14. Prairie soils
	15. Wiesenboden
4. Mildly leached soils	16. Solonetz
	17. Solodised solonetz and solodic soils
	18. Soloths
	19. Solonised brown soils
	20. Red-brown earths
	21. Non calcic brown soils
	22. Chocolate soils
	23. Brown earths
5. Soils with predominantly sesquioxidic clay minerals	24. Calcareous red earths
	25. Red earths
	26. Yellow earths
	27. Terra rossa soils
	28. Euchrozems
	29. Xanthothems
	30. Krasnozems
6. Mildly to strongly acid and highly differentiated	31. Grey-brown podzolic soils
	32. Red podzolic soils
	33. Yellow podzolic soils
	34. Brown podzolic soils
	35. Lateritic podzolic soils
	36. Gleyed podzolic soils
	37. Podzols
	38. Humus podzols
	39. Peaty podzols
7. Dominated by organic matter	40. Alpine humus soils
	41. Humic gleys
	42. Neutral to alkaline peats
	43. Acid peats

Figure 6.4: The soil classification system of Stace, Hubble and others, 1968. (The authors claim that the Great Soil Groups — those in the right-hand column — are arranged in order of degree of profile development and degree of leaching. Categories in the left-hand column presumably approximate Stephens' soil suborders.)

soils could not be shown on maps of continental extent. Soil families or soil series are frequently grouped on maps into *soil associations* which include whole catenas (see Chapter 5) as single mapping units. Associations can, however, be considered at the Great Soil Group level as well as at the family or series level. On very large scale soil maps, an additional level of classification is frequently needed: this is the *soil phase* which shows variations in the slope of the land, the degree of erosion of the soil and the effects of cultivation. Phase variations are more related to site than to the soil itself, but are quite important on maps for soil conservation.

STACE, HUBBLE AND OTHERS' SYSTEM (1968)

Stace, Hubble and others in *A Handbook of Australian Soils*, which appears in lieu of a fourth edition of Stephens' *Manual of Australian Soils*, present a soil classification which bears very little resemblance to Stephens' system. They discard solum classes and soil orders, the former being incorporated into their suborders and the latter not being used at all, as their soil orders include both pedocals and pedalfers. Their system of classification is shown in Fig. 6.4. The authors claim: 'the groups are now given in such an order that they represent overall a progressive increase in the degree of profile development and degree of leaching. Within the sequence a number of categories are recognised, beginning with those in which there is little or no profile development and ending with the soils dominated by the accumulation of organic matter. Thus the dark soils follow the category with minimal profile development, and are followed in turn by categories which are progressively more strongly leached and highly differentiated.' —Stace, Hubble and others, 1968.

One wonders why soils dominated by the accumulation of organic matter are regarded as more strongly differentiated and strongly leached than either the podzolised, or sesquioxide dominated soils, especially since the category includes neutral to alkaline peats which are scarcely leached or differentiated at all. Perhaps since this classification system merely represents a means for arranging a vast range of descriptive and analytical material in some order in a book, one should not be too critical of the scheme itself.

NORTHCOTE'S SYSTEM (1965)

Partly because of the problems of having to describe modal profiles and of intergrading of Great Soil Groups, Northcote's soil classification system nominally sets aside the Great Soil Group concept and aims at more precise definitions of the soil profile. His scheme is based on Australian soils, with an initial coding of 500 soils. His classification uses only the field properties of soil and so has a wider application than classifications like Guy Smith's, which rely partly on laboratory analysis of soils. The classification is designed to cater for surface soils, though buried soils, if sufficient of them remain, may also be coded.

Northcote's scheme uses letters and numbers, rather than names, to code soils. It is not necessary to carry the coding through to its completion when not enough information is available. The highest level of classification, *primary profile forms* (also called divisions), is made on the basis of the field determination of soil texture changes down the profile. Precise determination of texture classes requires practice, and best results are obtained if several pedologists compare their 'texturing'. Small differences in the texture classes down the profile could place soils in a different primary profile form. Compared with the other properties used for major divisions in Northcote's system, texture is perhaps the least objective. Four primary profile forms are recognised: three are based on texture alone, the fourth on the quantity of organic matter in relation to texture. The four primary profile forms are:

O soils dominated by organic matter, with the upper 12 inches containing 20% or more organic matter, provided that the clay content is 15% or less; or 30% or more organic matter, provided that the clay content is more than 15%.

U uniform primary profile forms: soils dominated by mineral matter but showing little textural change down the profile; that is, with no clearly defined textural boundaries within the solum and all the horizons included within one field-texture class.

G gradational primary profile forms: soils dominated by mineral matter and becoming increasingly finer-textured down the profile; the texture difference between consecutive horizons is not more than 1 to 1½ field-texture classes, but the range of texture through the entire solum exceeds one field-texture class.

D duplex primary profile forms: soils dominated by mineral matter and showing marked texture contrast between the eluvial and illuvial horizons of at least 1½ to 2 field-texture classes; the *B* horizons are clayey and their upper boundary with the eluvial horizons is usually sharp.

Northcote recognises six field-texture classes as follows:

1. Sands: sand, clayey sand, loamy sand 2. Sandy loams: sandy loam to light sandy clay loam	The coarse textures

3. Loams: loam, fine sandy loam, silty loam, sandy clay loam 4. Clay loams: fine sandy clay loam, silty clay loam, clay loam	The medium textures
5. Light clays: sandy clay, silty clay, light clay 6. Medium-heavy clays: medium clay, heavy clay	The fine textures

Note that Northcote uses texture classes in addition to those listed in Chapter 2 and shown in the texture triangle (Fig. 2.7).

Northcote's field-texture classes include several single texture classes, so a soil can still be classified as uniformly textured if its horizons are included in two classes in Fig. 2.7. The following profiles are classified as uniform, the textures going from top to bottom:

loam, fine sandy loam, sandy clay loam;
sandy clay, silty clay, light clay.

The following profiles are classified as gradational:

sand, loamy sand, sandy loam;
loam, clay loam, light clay;
clay loam, silty clay, medium clay.

In duplex profiles, marked texture changes are required so the subsoils are usually clayey in texture. In addition to marked change in texture, the change must occur over less than 4 inches, i.e., the boundary between the light and heavy textured horizons must be sharp. The following profiles are classified as duplex:

sand over clay; sand over sandy clay loam;
loam over clay; loam or clay loam over medium or heavy clay.

Northcote's *subdivisions* are not based on a common property, but on those properties which are most useful in each primary profile form. Organic primary profile forms are not subdivided, and three different properties are used for each of the three remaining profile forms. Uniform primary profile forms are subdivided on the basis of texture, and on whether the fine textured soils crack. There are four subdivisions in the U primary profile forms:

Uc coarse textured

Um medium textured

Uf fine textured, non-cracking

Ug fine textured, cracking.

There are two subdivisions within *G* primary profile forms depending on the presence of lime:

Gc	profiles are calcareous throughout
Gn	profiles are not calcareous throughout, but lime may be present in the *B*, *C* or *R* horizons.

The *D* primary profile forms are subdivided into five parts depending on the Munsell colour rating of the top six inches of the clayey *B* horizon:

Dr	red *B* horizon (Munsell hue red or redder than 5YR)
Db	brown *B* horizon (Munsell hue yellower than 5YR and value/chroma rating of 5)
Dy	yellow *B* horizon (Munsell hue yellower than 5YR and value/chroma of 2 or 4)
Dd	dark *B* horizon (any Munsell hue, with value/chroma of 1)
Dg	pale or gley *B* horizon (any colour on the Munsell gley chart or any hue with value/chroma rating of 3).

All the soils in the *D* texture may be mottled or whole coloured (Northcote considers a soil as mottled if more than 10% of its matter is differently coloured).

Primary profile forms (divisions) and subdivisions are shown in Fig. 6.5.

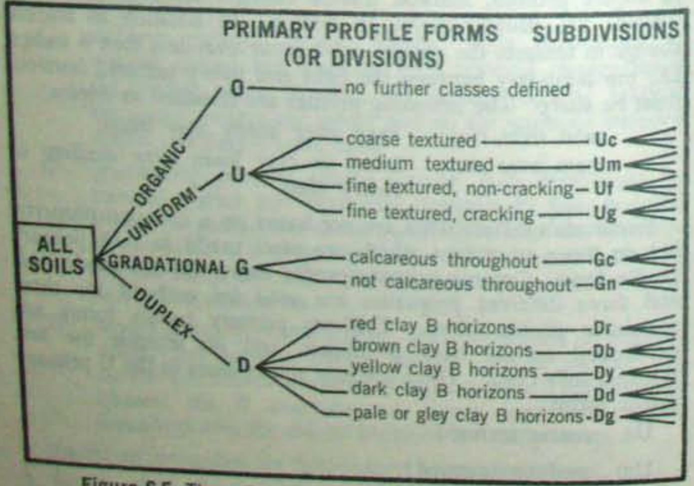

Figure 6.5: The primary profile forms (divisions) and subdivisions used in Northcote's Soil Classification (1965).

Northcote's sections, subsections, classes, subclasses and principal profile forms also depend on the properties found most useful, rather than properties common to all categories. Numbers are allotted for *sections*, *classes* and *principal profile forms*, while *subsections* and *subclasses* represent groupings of these numbers and are allotted letters, e.g.,

subsection Uc1 a includes Uc1.1 to Uc1.2
subsection Uc1 b includes Uc1.3 to Uc1.4

Sections are designated by the first number, for example, the number 1 in Uc 1; classes are designated by the second number, for example, the 2 in Uc1.2; and principal profile forms are designated by the third number, for example, the 3 in Uc1.23. The main properties used in sections, classes and principal profile forms are the degree of horizon differentiation, colour, the nature of the *E* horizon, the cementation of *A* and *E* horizons, the structure of the *B* horizons, the presence of pans, especially carbonate ones, and the soil reaction trend which brings out the general pattern of *pH* change down the profile. The *pH* is measured with a *pH* meter or indicators which correlate with meter measurements, such as Raupach's Indicator widely used in Australia to estimate soil reaction. Fig. 6.6 shows the sections, classes and principal profile forms within the subdivision Uc to illustrate the various properties used at each level of classification.

At its present stage of development, Northcote's system of soil classification is most useful in soil mapping. The pedologist need merely code the soil in the field by the appropriate letters and numerals, rather than write a lengthy description of the soil profile at each site. The compilation of the sheets of the Atlas of Australian Soils at 1:2,000,000 was started in 1960, mainly by Northcote. The soils are listed in the main legend in terms of the Northcote code letters; however, explanatory booklets with each map provide verbal descriptions of the code letters and list Great Soil Group equivalents to the various principal profile forms. These maps provide the most up-to-date coverage of Australian soils and, together with Stace, Hubble and others' (1968) map at the back of the *Handbook of Australian Soils*, can usefully guide any soils excursions. However, students need to be aware of the small scales of these maps, and the possibility of encountering intrazonal soils not shown on them.

GUY SMITH'S SEVENTH APPROXIMATION (1960)

Guy Smith's Seventh Approximation is also being revised, but it is hoped that new soils will be accommodated within the framework of the 1960 edition of the classification. This classification is also based on soil properties, but, unlike Northcote's

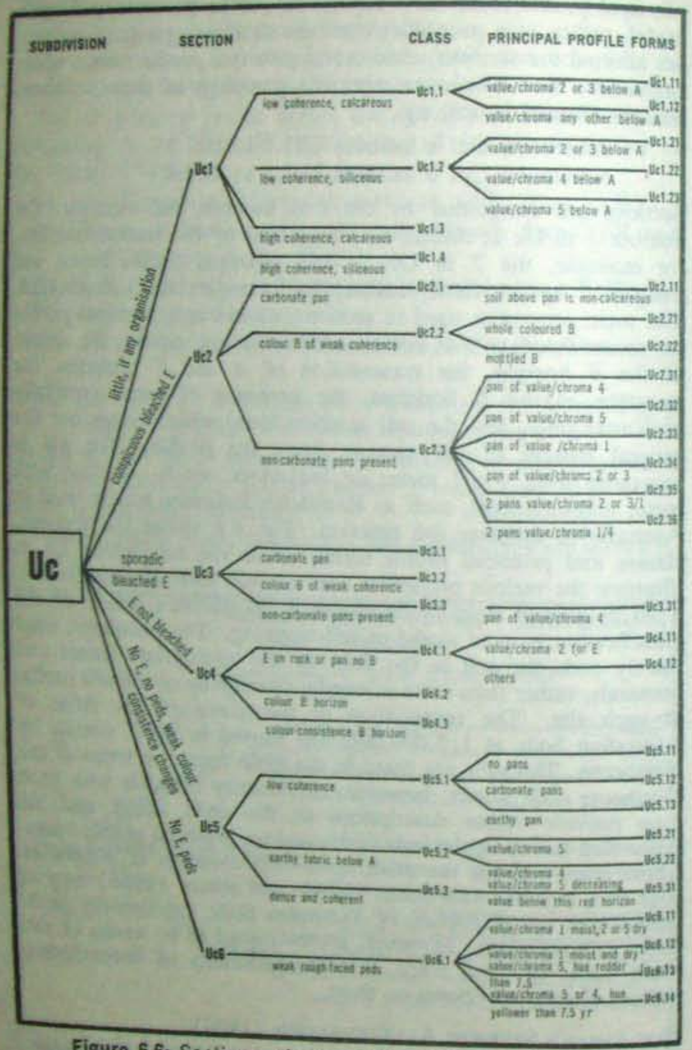

Figure 6.6: Sections, classes and principal profile forms within the Uc subdivision of Northcote's Classification (1965).

system, laboratory measurement is needed for full classification, though some of the properties are observable in the field. Preference is given to properties which are estimated quantitatively rather than qualitatively. The properties selected are those which relate to the soil-forming processes and indicate soil genesis. Where properties have equal genetic significance, preference is given to the one that has most influence on plant growth. Like Northcote's classification, the division of the categories is not made on the basis of common properties, but according to properties most useful to that division. Because American soils, on which this classification is based, tend to have their upper horizons altered by man (more so than Australian soils), emphasis is placed on the subsoil.

There are ten soil *orders* in the system. These orders correspond quite closely to the soil types of the early Russian pedologist, Dokuchaev. The orders are based on the properties of soils that reflect similar processes of horizon formation. They include zonal, intrazonal and azonal soils and pedalfers and pedocals and perhaps parallel Stephens' suborders rather than his orders or solum classes, though it is difficult to compare the two systems. The soil orders, with a brief summary of their main features, are as follows:

1. Entisols (from *ent* meaning recent) correspond to azonal soils
2. Vertisols (from *verto* meaning turn), the cracking, strongly gilgaied clayey soils
3. Inceptisols (from *inceptum* meaning beginning), young undifferentiated soils including brown forest soils and humic gley soils
4. Aridisols (from *aridus* meaning dry), the soils of the deserts, including solonchaks, brown desert soils, red desert soils
5. Mollisols (from *mollis* meaning soft), the soils of the semi-desert, including solonetz, prairie soils, chernozems and rendzinas
6. Spodosols (from *spodos* meaning wood or ash), podzols, brown podzolics and groundwater podzols
7. Alfisols (from *alf* meaning pedalfer), grey-brown podzolics, grey wooded soils and non-calcic brown earths
8. Ultisols (from *ultimus* meaning last), red-yellow podzolics and reddish lateritic soils
9. Oxisols (from *oxide* meaning a compound with oxygen), lateritic soils, red earths
10. Histosols (from *histos* meaning tissue), hydromorphic organic soils such as bog soils.

Suborders are based on the characteristics producing the greatest degree of genetic homogeneity. The colour of the wet soil is used with other properties. The suborders form prefixes to the orders; for example, a strongly mottled entisol is called an aquent. Some of the suborder prefixes are:

acr (from *acros* meaning highest)—most strongly weathered
alb (from *albus* meaning white)—presence of a bleached horizon
aqu (from *aqua* meaning water)—characteristics associated with wetness, i.e., mottling
arg (from *argilla* meaning clay)—a horizon with illuvial clay
ferr (from *ferrum* meaning iron)—the presence of iron
psamm (from *psammos* meaning sand)—sandy textures
ust (from *ustus* meaning burnt)—of dry climates, with hot summer.

The *great groups* are defined mainly by the presence in them of diagnostic horizons and the arrangement of those horizons. Some suborders have only one great group. The great groups, which number more than 40, form prefixes on the suborders and include all the prefixes used in the suborders as well as new prefixes. Examples of the additional prefixes are:

brun—dark colours
cru—crusting
cry—cold
crypt—hidden
gloss—tongued
therm—warm.

Subgroups can be defined only with reference to the great groups and refer mainly to the intergrades between the great groups. An adjective is used for the subgroups, e.g., *orthic* which refers to the normal great group. Some of the adjectives are formed from the great group prefixes, but others such as ruptic, meaning broken, are used as well.

Categories from *family* down rely on local names as in most other classifications, but emphasise the properties important to plant growth. *Series* differences generally relate to the original name of the feature.

The essential starting point for classification according to the Seventh Approximation is the *pedon* which is the smallest volume that can be called a soil in its own right. It has three dimensions and is defined by Guy Smith as 'the full solum to within the underlying unconsolidated rock, whose lateral cross section is hexagonal or circular in shape and from one to ten square metres in size'. The pedon forms the soil *phase* and from here the pedologist works up to the other levels of classification. The lower levels in the classification rely most heavily on laboratory determinations.

The Seventh Approximation has been regarded as cumbersome by some observers, and resistance to it has been strong, especially in Australia. Australian pedologists tend to prefer a simpler classification which has application to field studies; and so far they have not produced maps based on the Seventh Approximation.

FAO/UNESCO Soil Units for Soil Map of the World

The soil classification used in the FAO/UNESCO system is designed to produce the world soil maps at a scale of 1:5,000,000, so lower members of the soil classification hierarchy are not needed. The soil mapping units correspond most closely to the Great Soil Group level, but the soils within each subdivision are not necessarily all at the same level of generalisation.

Because the classification is an international one, the soil names selected were ones that did not change much on translation and had similar meanings in different countries. Where possible, traditional soil names are used, for example, podzols, chernozems, solonetz, rendzinas and lithosols, but names which have become recently popular in some of the newer soil classifications are also included, for example, vertisol from the Seventh Approximation and gleysol from the Canadian classification. Where soil terms are particularly confusing in soils literature, such as podzolised and brown forest soil, new terms have been selected in an attempt to avoid the confusions.

The names for the soils are taken from many different soil classifications from various countries, including Australia (Stephens' classification, 1962, in *A Manual of Australian Soils* is used). The soil names are derived therefore from Russian, Latin, Greek, Austrian, Japanese and other names. Fig. 6.7 lists the soils units and their abbreviations used on the maps. These abbreviations are either the first letter of the soil name, for example, *C* for chernozem, or the first letter of a soil name similar to the one used, for example *B* for cambisol which means brown forest soils. Not enough single letters were available, so lithosols (*I*), paramosols (*U*) and rendzinas (*Y*) had to be given arbitrary abbreviations.

Fig. 6.7 shows that the FAO/UNESCO system recognises 79 Great Soil Groups. The world soil maps being produced on the basis of this classification will be the only ones available in such detail. They cover every continent and will provide an invaluable reference source for all branches of geography and other sciences, as well as pedology. Many libraries will undoubtedly purchase these maps, so an explanation of the soils names used will be useful.

The main groupings of soils (fluvisols, rhegosols, etc.) may be considered as soil suborders. Many of the terms used in these suborders are repeated in the Great Soil Group names, so though some of the words may be new, the terminology is not as

Soil Name	Abbreviation
FLUVISOLS	J
Dystric fluvisols	Jd
Eutric fluvisols	Je
RHEGOSOLS	R
Dystric rhegosols	Rd
Eutric rhegosols	Re
ARENOSOLS	Z
GLEYSOLS	G
Ochric gleysols	Go
Fluvic gleysols	Gj
Humic gleysols	Gh
Calcic gleysols	Gk
Salic gleysols	Gs
Sodic gleysols	Gn
Thionic gleysols	Gd
Plinthic gleysols	Gp
Histic gleysols	Gm
RENDZINAS	Y
RANKERS	Q
ANDOSOLS	T
Halpic andosols	Tn
Vitric andosols	Tv
Gleyic andosols	Tg
VERTISOLS	V
Pellic vertisols	Vp
Chromic vertisols	Vc
ERMOSOLS	E
Halpic ermosols	En
Luvic ermosols	El
Calxic ermosols	Ek
Gypsic ermosols	Es
XEROSOLS	X
Halpic xerosols	Xn
Luvic xerosols	Xl
Calxic xerosols	Xk
Gypsic xerosols	Xs
HALOSOLS	S
Ochric solonetz	So
Humic solonetz	Sl
Gleyic solonetz	Sg
Ochric solonchak	Sa
Humic solonchak	Sh
Takyric solonchak	St
Solod	Sw
PLANOSOLS	W
Ochric planosols	Wo
Humic planosols	Wh
CASTANOZEMS	K
Halpic castanozems	Kn
Luvic castanozems	Kl
Calcic castanozems	Kk
CHERNOZEMS	C
Halpic chernozems	Cn
Luvic chernozems	Cl
Calcic chernozems	Ck
PHAEOZEMS	H
Halpic phaeozems	Hn
Luvic phaeozems	Hl
Gleyic phaeozems	Hg
CAMBISOLS	B
Ochric cambisols	Bo
Humic cambisols	Bh
Eutric cambisols	Be
Calcic cambisols	Bk
Vertic cambisols	Bv
LUVISOLS	L
Brunic luvisols	Lb
Chromic luvisols	Lc
Ferric luvisols	Lf
Rhodic luvisols	Lr
Albic luvisols	La
Glossic luvisols	Ll
Plinthic luvisols	Lp
Gleyic luvisols	Lg
ACRISOLS	A
Helvic acrisols	Ay
Rhodic acrisols	Ar
Humic acrisols	Ah
Plinthic acrisols	Ap
Gleyic acrisols	Ag
PARAMOSOLS	U
PODZOLS	P
Humo-ferric podzols	Po
Ferric podzols	Pf
Humic podzols	Ph
Placic podzols	Pi
Gleyic podzols	Pg
FERRALSOLS	F
Helvic ferralsols	Fy
Ochric ferralsols	Fo
Rhodic ferralsols	Fr
Humic ferralsols	Fh
Plinthic ferralsols	Fp
HISTOSOLS	M
Drystic histosols	Md
Eutric histosols	Me
LITHOSOLS	I

Figure 6.7: FAO/UNESCO soil units used in the world soil map project (1968).

formidable as it first appears, especially since many of the new terms are derived from well-known Latin roots such as fluvisol from *fluvius*, meaning a river, and arenosol from *arena*, meaning sand. The following notes give the derivation of the suborder names and a very brief description of the soils.

Fluvisols (Latin *fluvius*, a river) alluvial soils with little profile differentiation

Rhegosols (Greek *rhegos*, blanket) very poorly developed soils overlying hard rock

Arenosols (Latin *arena*, sand) poorly differentiated sands

Gleysols (Russian *gley*, mucky soil mass) soils containing reduced or mottled horizons resulting from excess water

Rendzinas (Polish word meaning shallow, sticky soils over limestone)

Rankers (Austrian *rank*, steep slope) shallow soils on siliceous material; no horizon formation

Andosols (Japanese *an*, dark) soils formed from material rich in volcanic glass and having a dark surface horizon

Vertisols (Latin *verto*, turn) heavy textured, deeply cracking soils prone to gilgai formation (churning)

Ermosols (Latin *eremus*, solitary, desolate) soils of the open deserts; desert sand plain soils and desert loams; poorly differentiated desert soils

Xerosols (Greek *xeros*, dry) desert soils with profile differentiation, but no strong salt accumulations

Halosols (Greek *hals*, salt) desert soils with salt accumulation and sodium domination; differentiated profiles

Planosols (Latin *planus*, flat, level) soils in level or depressed topography with poor drainage

Castanozems (Latin *castaneo*, chestnut; Russian *zemlya*, earth) soils rich in organic matter with a brown or chestnut tinge at the surface

Chernozems (Russian *chern*, black; *zemlya*, earth) soils rich in organic matter with a black colour at the surface

Phaeozems (Greek *phaios*, dusky; Russian *zemlya*, earth) soils rich in organic matter having a dark coloured surface

Cambisols (Latin *cambiare*, change) soils in which changes in colour, structure and consistence have occurred as a result of weathering *in situ* rather than illuviation

Luvisols (Latin *luvi*, perfect tense of *luo*, to wash) soils in which clay illuviation has occurred, but not strong base depletion

Acrisols (Latin *acris*, very acid) soils with illuvial clay and strong base depletion

Paramosols (Spanish *paramo*, land with scarce vegetation) soils of high altitude grasslands

Podzols (Russian *pod*, under; *zola*, ash) soils with a strongly bleached eluvial horizon

Ferralisols (Latin *ferrum*, iron; and *al* from aluminium) soils high in sesquioxides

Histosols (Greek *histos*, tissue) soils rich in fresh or partly decomposed organic matter

Lithosols (Greek *lithos*, stone) soils with hard rock at a very shallow depth.

The following notes give the derivatives of the Great Soil Group prefixes not already listed under the suborders above:

Dystric (Greek *dys*, ill, infertile) soils with a very low *pH*

Eutric (Greek *eu*, good, fertile) soils with a higher *pH* from recent deposits

Ochric (Greek *ochros*, pale) soils with a light coloured surface soil

Humic (Latin *humus*, earth) soils with surface layers rich in organic matter

Calcic from calcium, soils accumulating lime (calcium carbonate) or gypsum (calcium sulphate)

Salic (Latin *sal*, salt) saline soils

Sodic (from sodium) soils accumulating sodium

Thionic (Greek *theion*, sulphur) soils containing sulphur, usually as sulphates

Plinthic (Greek *plinthos*, brick) soils containing mottled clay materials which harden irreversibly on exposure

Halpic (Greek *halpos*, simple) soils with a simple, normal horizon sequence for their suborder

Vitric (Latin *vitrum*, glass) soils rich in glassy material

Pellic (Greek *pellos*, dusky, lacking colour) soils with low colour intensity

Chromic (Greek *chromos*, colour) soils with high colour intensity

Calxic (Latin *calxic*, lime) soils accumulating lime

Gypsic (from *gypsum*) soils accumulating gypsum

Takyric (Uzbek *takyr*, barren plain) soils of barren regions

Brunic (Anglo-Saxon *brun*, brown) brown coloured soils

Rhodic (Greek *rhodon*, rose) soils with a red to deep red colour

Albic (Latin *albus*, white) soils with a prominent bleached eluvial horizon

Glossic (Greek *glossa*, tongue) soils with tongues of eluvial material penetrating to the *B* horizon

Helvic (Latin *helvus*, yellow to yellowish-red) soils with reddish-yellow colours

Placic (Greek *plax*, flat stone) soils with a thin iron pan.

The FAO/UNESCO system recognises many more Great Soil Groups than do the classifications of either Stephens (1962) or Stace, Hubble and others (1968); this is understandable, for Australian soils contain only a small range of the world's soil spectrum. The presence of relict soils widens this range considerably, but, even so, soils of cold-wet and cold-dry regions are poorly represented in Australia.

By comparing the five soil classifications discussed in this chapter, one can see that the view expressed in the quotation at the head of this chapter is amply supported. The classifications designed for the purpose of mapping soils (for example, the FAO/UNESCO classification) and those which, although intended for wider purposes, are used mainly for mapping soils (for example, Northcote's classification) concentrate on field soil properties so that it is relatively easy to classify the soil to the level required. More comprehensive classifications which are designed for, or have found, wider application (for example, Guy Smith's) rely heavily on laboratory measurements of soil properties, and it is not so easy to fully classify a soil using these schemes. Notice, however, that all the classifications discussed in this chapter stress soil-forming processes, that is, the movement of constituents through the soil, causing horizon differentiation. The next important feature of these classifications is soil morphology, while the soil-forming factors receive little attention until the lower levels of classification are reached (type, series and phase). In this respect, these classifications differ markedly from early soil genetic classifications such as Dokuchaev's.

To discuss all the soils in the five classifications dealt with in this chapter would be beyond the scope of this book which is an introduction to soil studies. The classifications do, however, indicate the complex nature of the study of world soil distribution, and the wide selection of soils which could be chosen to illustrate the main soil-forming processes. Students who require more information on soils mentioned in the five classifications but not discussed in this book may refer to the works listed at the end of this chapter.

CLASSIFICATION SYSTEM USED IN THIS BOOK

The remaining chapters of this book will provide a simpler approach to soil classification than approaches thus far presented. The book stresses the dynamic approach to soils, emphasising the soil-forming processes which lead to the soil properties seen in the field. Its system of classification will be found to be based on *the main soil-forming processes*—the ones which are obvious in the field soil properties listed in Chapter 2.

These main soil-forming processes are, in other words, those that show most clearly what happens to the main products of the decomposition of mineral and organic matter in soils. The products of mineral weathering most conspicuous in soil properties are the alkali cations (especially sodium and potassium), the alkaline earth cations (especially calcium), iron and aluminium cations and the silicate clays. The alkali cations may combine with anions (especially chloride) in the soil to form salts, which not only appear as white particles within the soil but hinder plant growth. If they occur as cations on the soil colloids, they are easily detected by high *pH* values (above 8.5). The alkaline earth cations, on the other hand, tend more to combine with carbonate in the soil to produce less alkaline conditions. Lime (calcium carbonate) and gypsum (calcium sulphate) may also appear as white salts within the soil, but lime reacts to dilute hydrochloric acid and so is easy to detect in the field. These salts do not raise the *pH* as much as the alkali cations do (the *pH* of lime is only 8.3). The alkali and alkaline earth cations together constitute the *bases* found in the soil. Other cations, such as ammonium, which are also referred to as bases are less important to soil morphological properties than the alkali and alkaline earth cations. A base-depleted soil, then, is one lacking these cations and so has low *pH* values.

Iron and aluminium, together with manganese and titanium which are less important, may combine with oxygen and water within the soil to form the sesquioxides. Their presence is detected by the reddish and yellowish colours imparted by the hydrated iron oxides and to a lesser extent by black colours imparted by manganese and titanium oxides.

The amount of silicate clay in a soil horizon may be roughly estimated by texture determinations, heavy textures denoting much clay, and light textures lack of clay. Changes in texture down the profile indicate changes in clay content for various reasons such as illuviation of clay and higher rates of clay formation in certain horizons, the former being probably more important.

By way of simplifying the study of the soil-forming processes, humus (the products of the decomposition of soil organic matter) needs to be divided into two categories: humus which is high in the mobile, fulvic acid component and is brownish rather than black in colour; and humus which is high in the less mobile, highly coloured, humic acid component and is black in colour. Soils high in the fulvic acid component usually have marked illuviation of constituents, including humus itself, because fulvic acid assists the movement of mineral constituents through the profile. The soils high in humic acid are usually black coloured, with maximum humus at the surface, and decreasing with depth.

All these weathering products may move freely through the soil, around the soil's immobile constituents, or soil skeleton, which consists mainly of the sand and gravel fractions and to a certain extent the silt fraction. The soil skeleton has far less influence on soil properties than the mobile constituents, but if the latter are lacking in certain horizons, the skeleton may be responsible for the pale colour (quartz sand grains are white if uncoated by either iron or humus), sandy texture, apedal single grain structure (clay is needed to form peds) and low *pH*. Thus the properties of the soil skeleton are most important in *E* horizons from which mobile constituents have been eluviated.

There are four things that can happen to the products of mineral and organic decomposition in soils, depending on the circumstances (the five factors of soil formation). First, these weathering products may be completely removed from the soil profile because leaching exceeds their rate of formation. Second, they may accumulate throughout the soil profile because their rate of formation exceeds removal by leaching. Third, they may be illuviated from the *A* or *E* to the *B* horizons, and accumulate in the *B* horizons. Fourth, the converse may occur and they may rise from lower to upper horizons where they may concentrate: this rise operates through the soil's capillary pores by surface tension (see Chapter 3). We have considered only six products of weathering (alkali cations, alkaline earth cations, iron and aluminium, silicate clays, humus high in fulvic acid and humus high in humic acid) and only four outcomes for these products. Theoretically, there are twenty-four possible simple processes of soil formation, starting with removal of alkali cations, and proceeding to accumulation of salts from alkali cations in all horizons, illuviation of alkali cations and salt accumulation in *B* horizons, rise of alkali cations and salt accumulation in upper horizons and removal of alkaline earth cations, and so on. Fortunately, not all the twenty-four combinations are likely to occur, so the number of simple soil-forming processes to be considered is somewhat less. The most likely combinations are presented in Fig. 6.8.

Outcome of the Mobile Soil Constituents	Mobile Decomposition Products from Mineral and Organic Weathering in Soils					
	Bases					
	Alkali cations (especially Na⁺ K⁺)	Alkaline earth cations (especially ca²⁺)	Iron and aluminium which form sesquioxides	Silicate clays	Humus high in humic acid	Humus high in fulvic acid
Removal from soil profile	X	X			X	X
Accumulation throughout soil profile	X	X	X	X	X	
Illuviation to B horizons	X	X	X	X		X
Rise to upper horizons	X	X	X			

X denotes the process is a common, well-recognised one: blank space denotes the process is unlikely or rare.

Figure 6.8: Simple soil-forming processes in relation to the mobile soil constituents which dominate soil properties.

Removal of silicate clays from the profile is not likely, since they move in suspension rather than solution. Similarly, iron and aluminium are not very soluble, and once deposited as sesquioxides are quite insoluble, so they are unlikely to be completely removed from the soil profile; unlike silicate clays, however, when they do move, they are ions in solution and not particles in suspension and so can move freely upwards through the soil's capillaries. But upward clay movement is not common, because of the larger size of colloids which must move in suspension compared with ions which move in solution: clay particles tend to block the soil's capillaries when moving up and prevent further upward movement. Some surface hardpans possibly form in this way.

Humus high in humic acid is unlikely to be illuviated or to rise through the profile, for it is immobile; but humus high in fulvic acid, being mobile, is not so likely to accumulate throughout the profile and tends to be illuviated. Since the source of fulvic acid is the surface leaf litter, it is not likely to rise through the profile: once it is deposited in the *B* horizons, it is fairly insoluble.

As can be seen, some of the weathering products are less soluble than others and are therefore less prone to removal. Also, the constituents formed when the cations combine with anions in the soil, vary in their solubility—sesquioxides are most unlikely to dissolve again, whereas sodium chloride is quite soluble. Because of this variation in resistance to leaching, some combinations of soil-forming processes are less compatible than others; for example, salt accumulation is not likely to operate with the illuviation or capillary rise of iron and aluminium, because the former requires very weak, and the latter very intense, leaching; base accumulation is not likely to operate with illuviation of humus or iron because the latter requires rather low *pH* values. This incompatibilty of certain of the simple soil-forming processes makes the problem far less formidable.

From the above considerations, it is apparent that certain combinations of the simple soil-forming processes are more likely than others to occur in soils. These well-recognised combinations of simple processes may be referred to as *formal* processes, and are given names such as *podzolisation* (denoting base depletion, surface accumulation of organic matter, illuviation of humus, iron or clay); *laterisation* (denoting accumulation of sesquioxides at the expense of all other constituents including humus, with rise of iron and aluminium to form indurated sesquioxides near the soil surface); and *solodisation* (salt accumulation followed by some of the processes of podzolisation). The precise number of these simple processes that should be included within each formal process differs among pedologists.

The formal processes of soil formation give rise to the grouping of soils into suborders. It is possible, also, to group the formal processes in order to reduce the number of suborders; for example, laterisation may be included with ferralisation to form a group of soils dominated by sesquioxides, and calcification and salinisation may be included with solodisation to form a group of soils dominated by base accumulation in some or all horizons. This is what has been done in the following chapters, with the intention of grouping the formal processes on the basis of the dominant simple process; for example, Chapter 7 refers to all soils dominated by sesquioxide accumulation, regardless of whether the sesquioxides accumulate throughout the profile or are concentrated in certain horizons; and Chapter 10 refers to all soils accumulating silicate clays throughout the profile, though some soils also accumulate bases, while the remainder are base-depleted. Some problems are encountered with transitional soils, and their inclusion in certain chapters could be queried, but the most common and best developed soils within each grouping are stressed, rather than the transitional soils. The following chapters include pedalfers and pedocals when accumulation of lime is less important than other processes. Azonal soils are not discussed, partly because they have been mentioned in earlier chapters, but mainly because the purpose of this book is to demonstrate the processes of soil formation, and azonal soils either lack these processes or the processes are minimally developed.

The following outline of the soils to be further studied in the next six chapters indicates the main characteristics differentiating them:

Chapter 7	Soils Dominated by Sesquioxides. These soils are characterised by base depletion, removal of humus because decomposition exceeds its rate of accumulation, and domination of the profile by sesquioxidic clays, as distinct from silicate clays.
Chapter 8	Podzolised Soils. These have marked profile differentiation and are characterised by base depletion, the accumulation of a surface mat of organic matter, the illuviation of humus, iron and clay and the development of *E* horizons.
Chapter 9	Soils Dominated by Organic Matter. These soils are characterised by humus accumulation throughout the profile, but some soils show concentrations in certain horizons. Illuviation of constituents is impeded by the excess of organic matter.
Chapter 10	Soils Dominated by Silicate Clays. These soils accumulate silicate clay throughout the profile so that illuviation of constituents is impeded; commonly, these soils are associated with heavy textured parent materials such as loess, basalt or limestone.
Chapter 11	Soils Dominated by Accumulations of Bases. In these soils the alkali and alkaline earth cations combine with anions to form salts somewhere within the profile.
Chapter 12	Relict Soils in Australia. The morphology of these soils is not entirely related to the soil-forming processes operating in them at present: the soil-forming processes have changed over time in response to changing environmental conditions.

Fig. 6.9 lists the Great Soil Groups discussed in each of the above chapters and relates them to the other soil classifications discussed in this chapter. Not every soil in every classification mentioned in this chapter is discussed in the chapters which follow. This will be noted especially of the Great Soil Groups in the Seventh Approximation and the FAO/UNESCO system, emphasis being given to soils well represented in Australia. The names used for the Great Soil Groups in the following chapters are generally the traditional names. They are not necessarily used in the same sense as in Stace, Hubble and others' *Handbook of Australian Soils* (1968), as the writer does not always agree with their use of Great Soil Groups, especially when it departs radically from traditional, overseas use. Stephens' (1962) use of the Great Soil Groups is more in keeping with the aims of this book, as Stephens, with his wide knowledge of overseas soil morphology, uses the terms more traditionally and so relates Australian soils to the world soil spectrum.

Many maps show soil distribution in Australia. For the distribution of Great Soil Groups, students are referred to the maps at the back of Stephens' *Soil Landscapes of Australia* (1961), and Stace, Hubble and others' *Handbook of Australian Soils* (1968); but much more detail is provided by Northcote's *Atlas of Australian Soils* (1962-1968). Most physical geography textbooks (for example Strahler's various texts such as *The Earth Sciences* or *Introduction to Physical Geography*) include a map of Great Soil Groups of the World with which the larger scale Australian maps of soil distribution can be compared.

	Name of Great Soil Group used in this Book	Name used by Stephens, 1962	Name used by Stace, Hubble and others, 1968	Classification used by Northcote, 1965	Classification used by Smith in 7th Approximation, 1960	Classification used by FAO/UNESCO World Soil Map, 1968
CHAPTER 7 **Soils Dominated by Sesquioxides**	lateritic soil				oxisol	ferralsol
	krasnozem	krasnozem	krasnozem and euchrozem	Gn3.11, Gn3.14 Gn4.11, Gn4.31, Gn4.34	rhodustalf palexerult	rhodic luvisol rhodic acrisol
	red earth	red earth	red earth	Gn2.11, Gn2.12, Gn2.14, Gn2.15		
	yellow earth	yellow earth	yellow earth	Gn2.21, Gn2.22, Gn2.24, Gn2.32 Gn2.34, Gn2.61, Gn2.62, Gn2.64, Gn2.74, Gn2.4	orthox	ochric ferralsol
CHAPTER 8 **Podzolised Soils**	podzol	podzol	podzol	Uc2.2, Uc2.3	orthod	podzol
	ground-water podzol	ground-water podzol	humus podzol	Uc2.3, Uc4.3	humod	humic podzol
	brown podzolic	brown podzolic	brown podzolic	Dy2.62, Dy3.61, Dy4.41	hapludoll	phaeozem
	brown forest soil				dystrochrept	cambisol
	red podzolic	red podzolic	red podzolic	Dr2.21	hapludult	helvic acrisol
	yellow podzolic	yellow podzolic	yellow podzolic	Dy2.21, Dy2.22, Dy3.21, Dy3.22 Dy5.61	haplustult haploxerult	
	grey-brown podzolic	grey-brown podzolic	grey-brown podzolic	Db1.42, Dr3.12, Dy3.22	hapludalf	brunic luvisol
	gley podzolic	gley podzolic	gleyed podzolic	Dy3.41 Dy5.41, Dy5.42	aquod	gleyic podzol
CHAPTER 9 **Soils Dominated by Organic Matter**	alpine humus soil	alpine humus soil	alpine humus soil	Um7.11, Um7.12	histosol	histosol
	acid moor peat	moor peat	acid peat			
	humic gley soil	acid swamp soil	humic gley soil	Uf6.4 Uf6.6 Ug4.11		
	fen soil	fen soil	neutral to alkaline peat			
	moor podzol peat	moor podzol peat	peaty podzol	Uc4.3	humod	humic podzol
CHAPTER 10 **Soils Dominated by Silicate Clay**	shallow red soil			Db3.12, Dr4.12		
	chocolate soil	chocolate soil	chocolate soil	Gn3.21, Gn3.22, Gn3.24, Gn3.42	haplustoll	halpic castanozem
	prairie soil	prairie soil	prairie soil	Gn4.41, Gn4.42, Um6.21, Uf6.22, Uf6.32		
	black earth	black earth	black earth* and chernozem	Ug5.1, Ug5.4	pelloxerert	pellic vertisol
	chernozem			Um6.11, Gn3.43	haploboroll	chernozem
	wiesenboden	wiesenboden	wiesenboden	Ug5.16	pellustert	pellic vertisol
	grey and brown soil of heavy texture	grey and brown soil of heavy texture	grey, brown and red clay	Ug5.2, Ug5.3, Ug5.5	chromoxerert torrert	chromic vertisol
	terra rossa	terra rossa	terra rossa	Uc6.13 Um6.24, Um6.33	rhodoxeralf	chromic luvisol
	rendzina	rendzina	rendzina	Uf6.11	rendoll	rendzina
	grey, brown and red calcareous desert soil	grey-brown and red calcareous desert soil	grey-brown and red calcareous soil	Um5.11, Um5.61, Um6.22	orthent	rhegosol
CHAPTER 11 **Soils Dominated by Base Accumulations**	red-brown earth	red-brown earth	red-brown earth	Db1.13, Db1.33, Dr2.13, Dr2.23, Dr2.32, Dr2.33	rhodoxeralf	chromic luvisol
	non-calcic brown earth	brown earth	non-calcic brown soil	Db1.22, Dr2.12, Dr2.22	hapludalf	brunic luvisol
	solonised brown soil	solonised brown soil	solonised brown soil	Gc1.1, Gc1.2, Gc2.1, Gc2.2, Gn1.13	calciorthid	calxic ermosol
	solonchak	solonchak	solonchak	Uf1.4, Uf6.5, Uf6.6, Um1.1, Um1.2	salorthid	solonchak
	solonetz	solonetz	solonetz	Dd1.43, Dy2.43, Dy3.43	natrixeroll	gleyic solonetz
	solodised solonetz	solodised solonetz	solodised solonetz and solodic	various Db, Dd, Dr and Dy	natrustalf	ochric and humic solonetz
	solodic	soloth	soloth	Dr2.41, Dy2.41, Dy3.41, Dy5.41, Dy5.81	argialboll	solod
	desert loam	desert loam	desert loam	Dr1.12, Dr1.13, Dr1.32, Dr1.33, Dr1.42, Dr1.43	typic aridisol	ermosol
	desert sand	desert sandhill	siliceous sand	Uc1.21, Uc1.22, Uc1.23, Uc5.1	quartzipsamment	arenosol
	sierozem (grey desert soil)				mollic aridisol	xerosol
CHAPTER 12 **Relict Soils in Australia**	lateritic krasnozem	lateritic krasnozem			orthox, torrox, ustox	rhodic ferralsol
	lateritic red earth	lateritic red earth				
	calcareous lateritic soil	calcareous lateritic soil	calcareous red earth	Gn2.13		
	lateritic podzolic	lateritic podzolic	lateritic podzolic	Dy3.61, Dy3.62, Dy3.81	haplustalf	ferric and plinthic luvisol
	stony desert soil	stony desert tableland soil				
	red and brown hardpan soil	red and brown hardpan soil		Um5.3		

*These soils are distinguished on the basis of texture by Stace, Hubble and others, 1968. †These codings are taken from the map at the back of Stace, Hubble and others, Handbook of Australian Soils, 1968. Blank spaces denote the lack of a reasonably equivalent soil.

Figure 6.9: **Relationship between the Great Soil Group names used in this book and other soil classifications**

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Chapter Seven

Soils Dominated by Sesquioxides

CONTENTS

THE SOIL SUBORDER DOMINATED BY SESQUIOXIDES IS characterised by severe leaching in which bases, humus, and some of the silica are removed from the profile, leaving a highly weathered clayey soil dominated by the hydrated oxides of iron and aluminium, with lesser amounts of the hydrated oxides of manganese and titanium. These four hydrated oxides constitute the *sesquioxides.* The hydrated iron oxides, being reddish or yellowish in colour, usually dominate soil morphology. Black patches within the soil profile are often attributable to hydrated manganese and titanium oxides. As the hydrated aluminium oxides are not coloured, their presence may be ascertained only by chemical analysis.

There are two main monogenetic Great Soil Groups within this suborder: one dependent on the solum being invaded by the water table during soil formation, the other on it being freely drained. Under freely drained conditions, intense leaching produces a deeply weathered profile, reddish, well-structured and high in clay throughout, with little profile differentiation aside from a slight brownish colour (from organic matter) in the upper few inches. This Great Soil Group is referred to by various names: *krasnozem* (the name used in this book) and tropical red loam. Red earths and yellow earths, which are related to krasnozems also form under freely drained conditions. If the water table occurs within the solum, an entirely different profile results. Since sesquioxide-dominated soils are found in hot, wet areas having sporadic dry seasons, the water table level fluctuates within the solum, causing marked translocations of constituents into horizons. Also, because the lower horizons are frequently saturated, yellowish colours are featured. These translocations of constituents result in a well-differentiated profile with distinct horizons within the deeply weathered solum. Such profiles are referred to as *lateritic soils.* American pedologists use the term *latosol* (Strahler, 1961) for all tropical deeply weathered reddish soils, whether or not marked horizon development is present. Since true lateritic profiles are poorly developed in the United States, the term latosol is used more often for krasnozems.

All possible gradations between the undifferentiated krasnozems and the well-differentiated lateritic profiles are known, so, in keeping with the concept of the Great Soil Group presented in Chapter 6, krasnozems and lateritic soils represent fully developed members of a continuum containing less well-developed soils. In addition to gradations with each other, lateritic soils and krasnozems form gradations with other Great Soil Groups as the environment changes. Polygenesis introduces further complications, as lateritic soils frequently serve as the parent material for later soils which form under less intense weathering conditions. Many combinations of Great Soil Groups are possible in polygenetic lateritic soils, depending on the nature of the later environment; for example, if it is dry, lime will accumulate in the polygenetic soil to form a calcareous lateritic soil; if it is wet, the old lateritic soil may be podzolised and the new Great Soil Group is called a podzolic lateritic soil. Polygenetic lateritic soils are discussed in Chapter 12. The two Great Soil Groups outlined below, lateritic soils and krasnozems, are not the only members of this soil suborder, but they are the most common, fully developed monogenetic soils.

LATERITIC SOILS

The term *laterite* (from the Latin *Later, lateris* a brick) was first used by the English explorer, Francis Buchanan (later Hamilton) in his journey through Mysore, Canara and Malabar in India in 1807. He used the name for material at Angadipuram, a site later visited by Stephens (1961) whose profile description is the basis for Fig. 7.1 which indicates the whole sequence of horizons at Angadipuram. Buchanan, however, did not use the term laterite for the whole sequence of horizons, but gave it only to the material which, because it conveniently hardens on exposure to the air, the Indians use for building. This building material is concentrated in the indurated horizons shown in Fig. 7.1. Buchanan (1807) described this material as: 'immense masses, without stratification, full of cavities and pores, . . . containing a very large quantity of iron in the form of red and yellow ochres.'

Although the word laterite is rather loosely used in pedology, in this book it is used in the original sense intended by Buchanan,

Location: Near Angadipuram, Kerala, India.	**Land-use:** Cleared for paddy rice cultivation.
Slope: flat.	**Drainage:** poor, water table occurs within the solum.
Parent Material and Underlying Rock: gneiss.	**Average annual rainfall:** more than 120".

Horizon	Approx. Depth	Lower boundary	Colour	Moisture	Texture	Structure	Special Features
A eluvial	0-2 feet	Sharp, level	dark red	moist	loam	weak crumb	lateritic gravel scattered throughout, but increasing towards bottom of horizon
Box laterite (indurated horizon)	2-9 feet	gradational irregular	red and yellow with some black	moist	gravelly clay	pisolitic	so hard it cannot be cut with a knife; cut with a pickaxe to form building blocks which harden on exposure
Bg, Ox mottled horizon	9-20 feet	gradational irregular	mottled, red, yellowish-brown and grey	wet	clay	massive	paler colours increase with depth; water table above this horizon in wet season
Cg pallid horizon	20-38 feet	sharp, even	bleached grey and white	wet	sandy clay	rock structure preserved	water table above this horizon in both dry and wet season
R	38 feet +	fresh gneiss					

Figure 7.1: Lateritic soil in India (modified from Fox, 1936, and Stephens, 1961).

namely, for the indurated horizon. The whole sequence of horizons may be referred to as a *laterite profile*.

It can be seen from Fig. 7.1 that the laterite profile is very deeply weathered. This profile is certainly not the deepest on record, for in the Australian Desert, certain deeply weathered relict soils, with some horizons similar to those at Angadipuram, are more than 200 feet deep. The great depth of weathering is caused by the high rainfall and hot temperatures which instigate very thorough and deep decay of the parent material. The soluble constituents are leached from the profile and the less soluble iron and aluminium oxides accumulate. The high intensity of chemical weathering causes almost complete destruction of organic matter, although lateritic soils are frequently found under tropical rainforest where there is an appreciable leaf litter. So brown and black colours do not feature in the profile and humic acids do not greatly lower the *pH* of the soil during weathering (cf. the podzolised soils). Thus, although lateritic weathering is very intense, the soil is not very acid, the *pH* of the horizons commonly ranging from 5 to 7. Under these conditions, silica is soluble and may be leached from the profile as silicic acid.

The horizons shown in Fig. 7.1 are quite distinct. The profile is a simplified composite of descriptions by both Stephens (1961) and Fox (1936) in South India, near Angadipuram; for such horizons to develop, the water table needs to fluctuate regularly in the wet and dry seasons. This is not always the case, and the occurrence of lateritic soils on pre-weathered parent materials further complicates the profile. The profile in Fig. 7.1 is therefore more simple than one would expect to find in nature. However, it demonstrates the effect of the fluctuating water table on the genesis of the various horizons.

The *A* horizon of lateritic soils is often much thinner than that shown in Fig. 7.1 and may even be absent. It has eluvial properties, as its texture is lighter than the horizons beneath. Although it is still red in colour, some of its sesquioxides have most likely been eluviated to the indurated horizon or laterite. Sometimes the eluvial horizon is slightly darkened to reddish-brown in the upper few inches by organic matter which is soon destroyed and is unable to be illuviated to lower horizons. Lateritic gravel is quite common in the eluvial horizon as some sesquioxides are sporadically deposited above the laterite.

The laterite or indurated horizon is quite variable in morphology. Its composition is partly conditioned by the parent material. If aluminium predominates over iron, the indurated horizon may be termed bauxite rather than laterite. Some

indurated horizons, on rocks containing appreciable manganese, may contain large percentages of manganese oxide (this is used as the ore manganite). Laterite represents a thorough concentration of the residual elements of the parent material after the more soluble constituents have been removed. This concentration, which is more thorough than in the other horizons of the profile, is shown in Fig. 7.2 which gives the chemical composition of laterite, bauxite and the parent dolerite from profiles in British Guiana, analysed by J. B. Harrison (1933).

	Laterite	**Bauxite**	**Parent Dolerite**
SiO_2 and quartz	0.76%	9.80%	52.00%
Al_2O_3	10.54	59.00	17.27
Fe_2O_3 and FeO	75.08	2.07	11.16
H_2O	9.60	28.06	0.35
TiO_2	3.91	0.70	0.53
MnO	nil	nil	0.05
MgO	tr	nil	6.95
CaO	0.02	nil	8.80
K_2O	nil	0.07	0.18
Na_2O	nil	0.05	2.81
P_2O_5	0.03	0.38	0.01
Total	99.94	100.13	100.13

Figure 7.2: Chemical composition of laterite, bauxite and parent dolerite from British Guiana (J. B. Harrison, 1933).

In Fig. 7.2, silica, the bases (magnesium, calcium, potassium and sodium), and, surprisingly, manganese have been lost from the parent dolerite, and the laterite and bauxite are composed mainly of aluminium, iron oxides and water. Much of this water is the water of hydration of the sesquioxides. Titanium oxide has accumulated in the laterite but not in the bauxite. In all the laterites and bauxites analysed by Harrison (1933) manganese failed to accumulate, perhaps because the parent rocks were inherently low in manganese. Phosphorus is slightly higher in the laterite and bauxite than in the parent dolerite, possibly because there has been fixation of some organic phosphorus by the excess iron and aluminium.

Since the laterite is permanently above the water table, even in the wet season, it is thought that sesquioxide enrichment is partly the result of ion migration through the fine capillary pores, from the lower, saturated clayey horizons of the profile. Since the eluvial layer is usually not well-developed, illuviation contributes far less to the enrichment than upward capillary migration. In the dry season, iron and aluminium ions, rising to the surface as the subsoil dries out, are oxidised to form the sesquioxides which are thereafter insoluble. As this process can continue indefinitely, it is not surprising that such a thorough concentration of sesquioxides occurs.

The mottled horizon is the zone of the fluctuating water table: in the dry season, it is below the mottled horizon, in the wet season it is at the top of the mottled horizon. Thus in the dry season the mottled horizon takes on some of the properties of the laterite above it and is enriched in reddish and yellowish iron oxides and the other sesquioxides. In the wet season, the properties of the pallid horizon beneath develop: with reducing conditions prevailing, drab grey and white colours develop.

The pallid horizon is often kaolinised bedrock, with little soil organisation apparent. Iron and aluminium have migrated upwards from the pallid horizon, which is below the water table even in the dry season. Where wells have been sunk through the pallid horizon, they may lower the water table to within this horizon, but this is not a natural condition. Some blue and green colours may develop in the pallid horizon from the reduced iron remaining.

The mottled horizon is higher in sesquioxides than the pallid horizon, which is usually quite high in silica, some of which forms the kaolinite, bleaching the parent rock. Kaolinite is one of the most highly weathered aluminosilicate clay minerals (see Chapter 3) and in the pallid horizon can form from the weathering *in situ* of the primary aluminosilicates, particularly the feldspars. In some pallid horizons, deposits of quartz-like minerals occur in lenses. This material, which some pedologists refer to as 'grey billy' may be silica illuviated from the upper horizons of the profile. Re-deposited silica is more a feature of lateritic profiles on silica-rich rocks such as granite, than silica-deficient rocks such as dolerite, although Harrison (1933) notes much more silica in the lowest horizons of the profiles he analysed. Fig. 7.3 indicates the chemical composition of laterite and mottled and pallid horizons on dolerite in British Guiana. Note the enrichment in Al_2O_3 in the pallid horizon, coming mainly from the clay mineral gibbsite.

The formal process giving rise to the lateritic profile as seen in Fig. 7.4 is referred to as *laterisation*. This process is made up of several single processes, some of which are unique to lateritic soils and some of which are common to other soils. The unique processes are those that cause the net loss of silica from the profile and the upward migration of large quantities of iron and aluminium ions to be oxidised into sesquioxides in the laterite. Although some of the silica may be translocated to the lower horizons of the lateritic profile, a great deal of it is completely leached. This silica is removed from the profile by the waters percolating through the soil in the wet season. Some of the silica may be detected by analysing seepage water in tunnels cut through lateritic profiles. The silica eventually finds its way to the oceans where it is deposited. Harrison (1933) analysed rain water, seepage water, river water and sea water associated with lateritic soils. He found that seepage water contained much more silica than rain water, that river water in laterised areas contained much more silica than river water elsewhere, and that sea water near the mouth of rivers issuing from laterised areas contained more silica than sea water eight miles offshore.

	Laterite	**Mottled Horizon**	**Pallid Subsoil**	**Dolerite**
Quartz	11.64	12.28	47.41	1.60
SiO_2	4.46	21.89	3.30	49.69
Al_2O_3	26.81	24.94	26.33	15.20
Fe_2O_3 and FeO	36.65	28.49	10.67	14.28
H_2O	18.86	11.29	11.28	0.30
TiO_2	1.13	0.38	0.67	1.00
MgO, CaO, K_2O, Na_2O	0.20	0.75	0.79	17.90
P_2O_5	tr	0.16	tr	0.01
Total	99.75	99.98	100.95	99.88

Figure 7.3: Chemical composition of laterite, mottled and pallid horizon and parent dolerite, in British Guiana (J. B. Harrison, 1933).

The conditions under which silica is leached and deposited are not fully understood. Not only is the *pH* of the soil solution important, but the nature of the other cations and anions exerts a strong influence on the mobility of silica. A lot of the silica leached from lateritic profiles remains on the land to silicify rocks within the laterised area, so that laterisation and silicification are frequently associated. In inland Australia, the deep mottled and pallid horizons remaining as relict soils are capped, not by laterite or bauxite, but by silcrete, a hard crust of almost pure silica, whose origin has not yet been fully explained. More is said about this silcrete in Chapter 12 which examines relict soils.

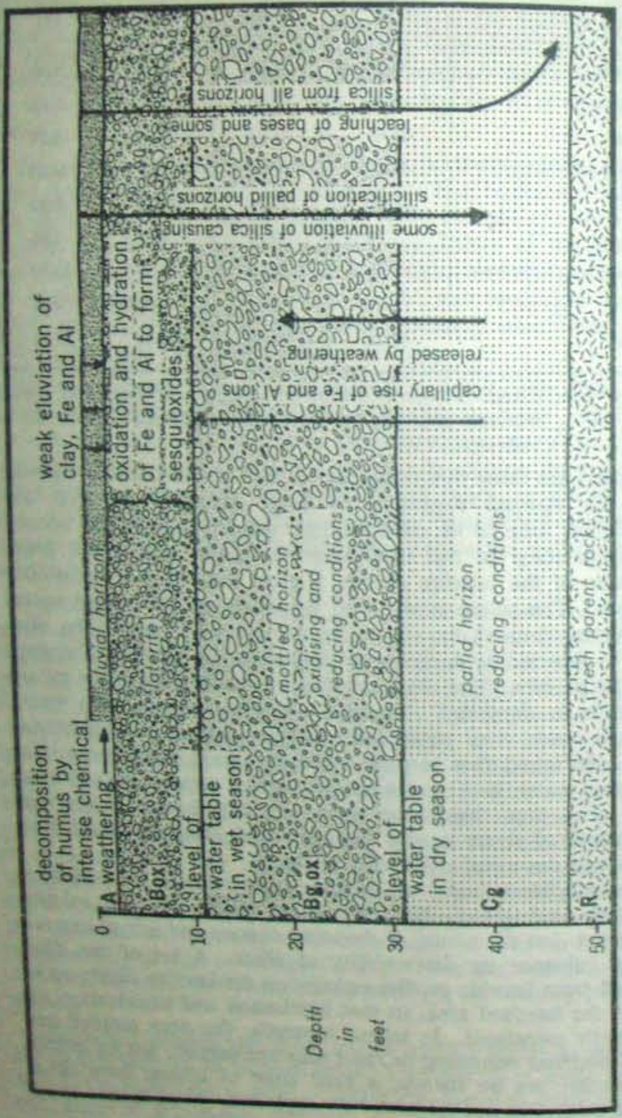

Figure 7.4: The process of laterisation.

Processes which form an integral part of, but are not unique to, laterisation are: base depletion, the thorough decomposition of humus, the high degree of decomposition of the parent material leaving a clayey soil lacking primary minerals, and the minimal illuviation of clay and sesquioxides. Base depletion is a feature of all soils found in areas where precipitation exceeds evaporation. Its morphological effects lie in the acid *pH* and lack of sufficient bases for prolonged cultivation. Shifting cultivation is necessitated in areas of lateritic soils, partly because of the low humus content, and partly because of this lack of mineral bases. Low fertility is also associated with the poor physical constitution of the soil (the laterite itself is rock-like in its hardness), and the fixation of mineral and organic phosphorus by the excess free iron and aluminium.

All the horizons of lateritic soils tend to be high in secondary minerals and to lack primary minerals, as, under such intense weathering, even quartz, one of the minerals most resistant to chemical weathering, is attacked. Under such intense weathering the more vulnerable minerals rapidly disappear, and from this disintegration comes the formation of the more stable clay minerals (usually from the kaolin group) and the sesquioxides, which lack silica. The processes of laterisation are shown diagrammatically in Fig. 7.4.

FACTORS AFFECTING LATERISATION

Lateritic soils are the most intensively weathered of all soils. Special environmental conditions are needed for laterisation, though the process is perhaps not as restricted as was previously believed.

The *climate* necessary for laterisation is one with very high temperatures throughout the year (usually the average monthly temperature exceeds 75°F in every month of the year), with a high average annual rainfall (usually more than 60″), and with a dry season to initiate upward migration of ions in evaporating soil water. The dry season need only be quite short, although long dry seasons as in savanna and monsoon climates are also conducive to laterisation, provided the rainfall in the shorter wet season is high enough. The high temperatures provide the necessary energy for laterisation, and the high rainfall provides the necessary moisture for deep chemical weathering. The climatic regions in which laterisation is known to be active today are equatorial (Koppen's *Af* climate), savanna (Koppen's *Aw* climate) and monsoon (Koppen's *Am* climate), but not all soils within these climatic regions are laterised: krasnozems are more widespread than lateritic soils.

Because lateritic soils are found within several tropical climatic regions, they support a wide variety of *vegetations*. Where the dry season is very short, tropical rainforest is associated with lateritic soils. Under intense weathering, the abundant leaf litter, the main source of plant nutrients, quickly decomposes and thus provides the nutrients essential for the trees, which adsorb some of the released cations and anions before they are completely leached from the soil profile. Removal of this rainforest to allow cropping swiftly eliminates the leaf litter, and soil productivity soon declines. Removal of the rainforest also hastens the drying of the laterite by free evaporation and so accelerates the development of its undesirable physical features. Clearing of lateritic soils also accelerates erosion of the meagre eluvial horizon, exposing the laterite at the surface and facilitating the hardening process. In monsoon climates, semi-deciduous monsoon woodland is associated with the lateritic soils; and in savanna climates, savanna woodland is thus associated. Drier savanna areas supporting mainly grassland contain relict, rather than actively forming laterite.

Microfaunal activity is another important aspect of the biotic environment of lateritic soils. Termites are considered by J. B. Harrison (1933) to be especially significant in maintaining aeration above the water table and therefore facilitating movement of water through the soil. Under the hot, wet conditions of laterisation both microbial and microfaunal activities are extremely high.

Laterisation is one of the few soil-forming processes which can operate on any *parent material*, from the most siliceous to the most basic. However, the nature of the parent material influences the morphology of the laterite profile, the rate of accumulation of the laterite, and even the chemical character of the laterite. Given similar durations of laterisation, siliceous rocks such as granite tend to form thinner, less enriched laterite, and deeper, better developed pallid horizons, in comparison with basic rocks such as dolerite in which the laterite and mottled horizons are well-developed and there is little bleaching in the pallid horizons. The nature of the parent material also conditions the structure of the laterite: on the Woronora Plateau, south of Sydney, pisolitic laterite is associated with shale, while massive and vesicular laterite, poorly enriched in sesquioxides, is associated with sandstone. The nature of laterisation, however, is similar on all parent materials, irrespective of the chemical character of these materials.

Because of the greater facility offered to water movement, pre-weathered parent materials such as alluvium or colluvium, form deeper laterite profiles than chemically similar rock parent materials. Enrichment of the laterite, and differentiation into horizons, are commonly very thorough on pre-weathered parent materials.

A great deal has been written on the *relief* factor in laterisation. Lateritic soils have frequently been associated with old erosion surfaces close to base level, called by the now rather obsolete term *peneplain* (e.g., Prescott and Pendleton, 1952; Whitehouse, 1941; and Woolnough, 1918). Relict lateritic soils, especially in Australia, have been associated with Tertiary peneplain surfaces which have since been uplifted and dissected. Different levels of laterite have generally been attributed either to differential uplift, or to more than one episode of laterisation separated by erosion. The association of laterite with peneplains is not wholly applicable to active tropical lateritic soils, which are found at several levels. However, some pedologists (e.g., Stephens, 1961) claim that these tropical laterites differ in age, and relate the highest level laterite to an old dissected peneplain surface. Since the presence of the water table within the solum is essential to the formation of laterite, peneplain conditions are undoubtedly suited to the laterisation process, where the climate also is suitable; but hot, wet situations on slopes may also provide suitable conditions, if the water table is within the solum. Thus active lateritic soils are not restricted to surfaces near base level, although such surfaces may be laterised. In Western Australia, it has been shown that relict laterite has formed on slopes up to 10°: these slopes are associated with the Darling Scarp which, being Jurassic in age, presumably pre-dates the Tertiary laterisation which formed these soils (Playford, 1954). Although Late Tertiary and post-Tertiary uplift may vertically separate some laterite profiles, others appear to have formed on surfaces that were initially quite steeply sloping, and laterite at different levels can no longer be taken as evidence of uplift, implying two episodes of laterisation. Certainly peneplain surfaces provide optimum relief conditions for laterisation, but active or relict lateritic soils are by no means restricted to such surfaces.

Not a great deal is known about the *time* required for the development of distinct lateritic horizons. Since inter-tropical areas have not been disturbed by glaciations, it is possible that some lateritic soils, like the tropical rainforests above them, have been forming continuously since the Tertiary. Their great depth could be the result of many thousands of years of undisturbed weathering, and many lateritic soils could be mature soils. However, this does not imply that active lateritic soils have been subjected to the same weathering environment since the Tertiary, as there is ample evidence for climatic change within, as well as outside, the tropics. Some lateritic soils within the tropics owe some of their morphological characteristics to weathering under previously more intensive conditions, although they are still undergoing desilication today. Kaye (1951) regards some of the lateritic soils of Puerto Rico more as palaeosols, although laterisation is still active there.

Outside the tropics, relict lateritic soils form many different later soils and it is here that the time factor has its greatest significance.

KRASNOZEMS AND RELATED SOILS

The word 'krasnozem' is derived from the Russian 'krasnaya zemlya' meaning 'red earth'. However, some pedologists distinguish between 'krasnozems' which are clayey textured and 'red earths' which are sandy, both soils being red throughout. Both soils arise from free drainage which promotes the formation of less hydrated iron oxides. Krasnozems are generally more important than red earths, especially within the tropics. Krasnozems develop in two sets of environmental conditions. First, they occur zonally within the hot, wet tropics on a wide variety of parent materials. Second, outside the tropics they occur intrazonally on parent materials high in iron and readily weathered minerals. Krasnozems formed in the latter conditions are usually shallower and have a higher *pH* than those developed in the former ones. Although temperatures are much lower, intrazonal krasnozems still require a high rainfall, perhaps more than 50" per annum, otherwise shallower red soils develop, and in even drier situations, prairie soils or chernozems (see Chapter 10).

Krasnozems are more common within the tropics than lateritic soils. This is fortunate since they are physically very well-endowed, although chemically, they still present serious problems. Lateritic soils and krasnozems are not the only inter-tropical soils, even in the wetter areas. Agriculturally, alluvium, alluvial regosols, and meadow soils on alluvium are just as important. In the drier parts of the tropics and at high altitudes, quite different soils develop.

Fig. 7.5 shows an intrazonal krasnozem on basalt on the South Coast of New South Wales. Similar soils are found in wet areas on basic rocks all along the Eastern Australian Coast. Along the Queensland Coast the type of parent material is less restricted, but heavy textured krasnozems are still not found on highly siliceous rocks such as sandstone.

Lack of profile differentiation is a feature of the krasnozem in Fig. 7.5. The soil is red throughout, with a slight browning caused

Location: Road cutting at Gerringong, South Coast, N.S.W.					**Land-use:** Sown pasture (mainly paspalum) for dairying.				
Drainage: well drained.					**Slope:** flat crest of hill.				
Parent Material and Underlying Rock: basalt.					**Average Annual Rainfall:** 52".				
Horizon	**Depth**	**Lower Boundary**	**Colour**	**Moisture**	**Texture**	**Structure**	**Root Penetration**	**pH**	**Special Features**
A_1 mulch	0"-½"	sharp, irregular	brownish-red	dry	clay loam	well developed fine crumb	strong root penetration to 4 feet	6.5	self-mulching layer
A_2	½"-8"	gradational, irregular	brownish-red	dry	clay loam	moderately developed ½" subangular blocky		5.5	
Bt	8"-8'	gradational, irregular	bright red, no mottles	dry	medium clay	well developed ½" subangular blocky		5.5	well developed clay skins; up to 4" diameter basalt stones scattered throughout
C	8'-15'	gradational, irregular	mottled red, yellow, grey and black	dry	stony clay	rock structure preserved		6.0	
R	15'+	fresh basalt, some weathering along columnar joints.							

Figure 7.5: Krasnozem at Gerringong, South Coast, New South Wales.

by humus in the upper few inches. No humus is evident in any other horizon—no *O* horizons are present and humus is not illuviated. As in the lateritic soils, humus decomposition is fairly thorough, though plant roots may penetrate quite deeply into the profile. Since the *A* horizon is clay loam and the *B* horizon is medium clay, there is some clay illuviation, but as the *B* horizon is so deep and uniformly textured, a great deal of the clay would presumably arise from weathering *in situ*. The high iron content, heavy texture, and domination of the clay by hydrogen result in a well-developed structure which further assists free internal drainage. Krasnozems are among the best-structured soils in the world. Like many krasnozems, the one shown in Fig. 7.5 is self-mulching, the friable *A* horizon forming a natural granular surface-covering, reducing evaporation losses from the profile.

In spite of the climatic high-leaching potential, the profile may be only slightly acid, because of the lack of humic acids (cf. podzolised soils) and the high base content of the parent material, though some krasnozems within the tropics do have *pH* values as low as 4.5. No free salts accumulate, and, except for the upper few inches where plants return cations to the surface, the clay is hydrogen-dominated, indicating that most of the bases have been removed from the profile.

The clay minerals in krasnozems are usually well-weathered: the kaolin group and hydrous oxides usually dominate. Not all krasnozems are as heavy textured as the one shown in Fig. 7.5 and some on the more resistant rocks contain appreciable quantities of sand, which, because of the high weathering intensity, consists of resistant minerals such as quartz and iron oxides. These soils grade into red earths in their properties. Krasnozems usually lack readily weathered primary minerals, such as olivine, augite, hornblende and feldspar.

Since sesquioxide enrichment is not as intense in krasnozems as in lateritic soils, desilication is not such an important process in krasnozem formation, and in some krasnozems it probably does not occur. Nicholls and Tucker (1956) found from 10% to 20% free iron oxide in krasnozems from Lismore, New South Wales. Compared with laterite, this is not high, but in terms of other soils, it is appreciable. The sesquioxides accumulate throughout the profile—they are neither illuviated (cf. podzolised soils) nor do they rise to the surface by capillary migration (cf. lateritic soils). The formal process giving rise to krasnozems is related to this accumulation throughout of sesquioxides. It is termed *ferralisation*, implying concentration of both iron and aluminium. Fig. 7.6 indicates the processes of ferralisation.

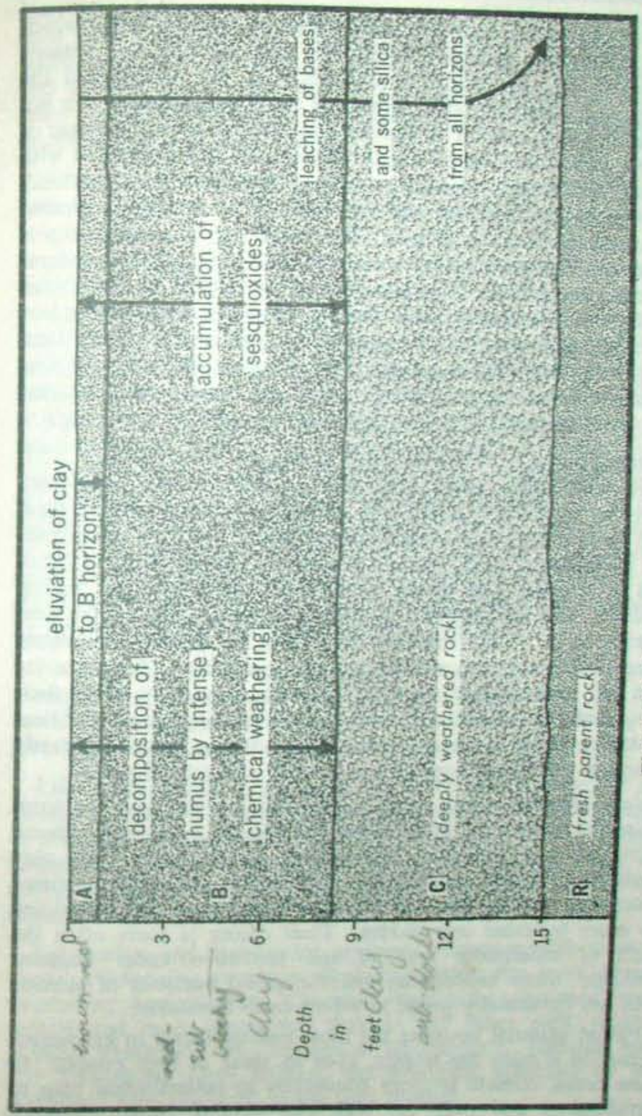

Figure 7.6: The processes of ferralisation.

Factors Affecting Ferralisation

Krasnozems are found in freely drained tropical areas which have both even rainfall distribution and distinct dry seasons (Koppen's *Af*, *Aw* and *Am* climates). A dry season and the presence of the water table within the zone of weathering are not critical to the formation of krasnozems, as they are in the case of lateritic soils. Like lateritic soils, krasnozems are associated with tropical rainforest, monsoon woodland and savanna woodland, and are widely used for shifting cultivation, presenting similar problems once they are cleared. Their ability to fixate phosphorus is just as high as that of lateritic soils. Cropped krasnozems gradually accumulate total phosphorus, for much of the added fertilizer phosphorus is soon fixated. Their structure does not deteriorate as a result of cropping, as lateritic soil structure does, though the high rainfalls make them susceptible to erosion. Tropical krasnozems are also found in areas too dry for lateritic soils, and this partly explains their more widespread occurrence.

Extra-tropical krasnozems (such as the one shown in Fig. 7.5), before cultivation, were frequently associated with rainforest. Their heavier texture, higher *pH*, and better structure, make them more productive than neighbouring highly leached podzolised soils on siliceous rocks such as sandstone. Although the krasnozems of Coastal New South Wales present phosphorus fixating problems, they are generally preferred for cultivation to the acid podzolic soils. They respond quite well to grazing as grasses return nutrients to the surface in large quantities. They are the main soil of the sugar coast of Australia, especially when their parent material is basalt. To maintain yields, nitrogenous fertilizer is usually added with superphosphate. Parent material greatly influences the morphology of krasnozems and related soils.

Very siliceous rocks, even within the tropics, tend to form *red earths* rather than krasnozems. Red earths are sandy textured throughout and, like most krasnozems, do not have strong clay illuviation. *Yellow earths* which are yellow and sandy textured throughout tend to form on siliceous parent materials containing the more hydrated iron oxides. Their colour is more often the result of inheritance than of soil formation under impeded drainage. Some sandstones, and the pallid horizons of lateritic soils, are particularly prone to yellow-earth formation.

Parent material becomes the most limiting factor in krasnozem formation outside the tropics, even in areas of high rainfall. In these areas, climate is more conductive to podzolisation than to ferralisation, and only the high clay-forming parent materials (e.g., basalt) resist it. However, not all such rocks in resisting podzolisation, form krasnozems. The chocolate shales of the Narrabeen Series, near Sydney, form stony chocolate earths which tend to be shallower than krasnozems, as well as differing in their colour and gravel content. Their uniform profile, weakly acid *pH*, and association with rainforest make them more akin to krasnozems than to podzolic soils. Ferralisation rather than podzolisation dominates on chocolate shales.

Krasnozems and lateritic soils are often in catenary association, both as active and relict soils. As active tropical soils, krasnozems frequently occupy the well-drained slopes, and lateritic soils form in the depressions where the water table invades the profile. As relict soils, very often the reverse occurs—eroded lateritic soils occupy dissected residuals of the old uplifted erosion surfaces, and krasnozems form downslope on the iron-rich colluvium. Lateritic colluvium is particularly prone to krasnozem formation, and many krasnozems owe their high iron content and great depth partly to pre-weathering. Krasnozems and lateritic soils are often intimately associated in relationships not quite as simple as the last two. Krasnozems often contain a laterite horizon without the development of either mottled or pallid horizons. It is often difficult to ascertain whether the laterite formed before, during, or after the krasnozem. Many relict lateritic soils are overlain by deeply weathered krasnozems above the laterite. Such krasnozems could have weathered from the old lateritic profile, or formed on an overlying deposit emplaced after laterisation.

Because they are found in similar areas, slight environmental changes can transform lateritic soils into krasnozems, and vice versa. Laterisation and ferralisation are similar processes, the former perhaps being a more extreme version of the latter.

Like lateritic soils, krasnozems are relict soils both within and outside the tropics. Relict krasnozems and red earths are common in Australia (Chapter 12) but these soils are actively forming under the present climate on many parent materials along the northern and northeastern coast, and further south on basic parent materials.

Laterisation and ferralisation represent extreme combinations of single soil-forming processes operating under the most intense weathering environment. Chemical weathering proceeds more rapidly and more thoroughly than in any other soils, making physical weathering, by comparison, unimportant. Lateritic soils and krasnozems form striking morphological contrasts to the podzolised soils discussed in the next chapter.

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Chapter Eight

Podzolised Soils

CONTENTS

THE WORD 'PODZOL' IS DERIVED FROM RUSSIAN WORDS MEANING 'ashes underneath' and was first used to describe soils with an ashen grey horizon immediately beneath the organic topsoil. This horizon, which is named the *E* horizon in the FAO/UNESCO system, is called the *bleicherde*. The name podzol was subsequently extended to include soils that were without bleicherde but similar in other properties to soils containing bleicherde. So although many podzolised soils do contain bleicherde, it is not an essential feature of all podzolised soils. The essential properties of podzolised soils are severe base depletion throughout the profile resulting in hydrogen-dominated clays, low *pH*, and marked horizon differentiation caused by illuviation. Although many podzolised soils contain a surface accumulation of acid organic matter (*O* horizons) soils without this may still be referred to as podzolic, especially in Australia.

Podzolisation is the name of the formal process which gives rise to podzolised soils. It is made up of several processes, many of which operate together. These processes are listed below; the first one, and at least one other, are necessary before a soil can be referred to as podzolised or podzolic:

1. Leaching of bases from the entire profile
2. Accumulation of a surface mat (*O* horizons) of acid organic matter
3. Eluviation of iron and humus from the bleicherde (*E* horizon)
4. Illuviation of humus
5. Illuviation of iron, often accompanied by illuviation of aluminium
6. Illuviation of clay.

The podzolised soils may be classified on the basis of these six processes. The term *podzol* is generally restricted to soils in which

Name of podzolised soil (alternative names are given in parentheses)	Dominating processes (X indicates process operates strongly)					
	1. Leaching of bases	2. Accumulation of surface mat of acid organic matter (O horizons)	3. Eluviation of iron and humus from the bleicherde (E horizon)	4. Illuviation of humus	5. Illuviation of iron and aluminium	6. Illuviation of clay
Podzol (nomopodzol)	X	X	X	X	X	may be minimal
Brown podzolic (crypto-podzolic)	X	X	absent	absent	hidden by humus accumulation	may be minimal
Red and yellow podzolics (amphipodzolics)	X	usually minimal	X	absent	X	X
Grey-brown podzolics (lepto-podzolics)	X	usually minimal	X	absent or minimal	absent or minimal	X

Figure 8.1: The podzolised soils classified according to dominating processes.

all six of the processes operate, though silicate clay illuviation is usually minimal. Soils in which the first process and only some of the remaining five processes operate are referred to as podzolics rather than podzols.

Podzols and podzolic soils are zonal soils and are associated with the forests of moist, temperate areas. They may also occur intrazonally, especially on very siliceous rocks. Intrazonal, hydromorphic podzols and podzolics form under impeded drainage which causes marked gleying. Many variations are possible in gleyed podzolised soils, depending on the number of podzolisation processes operating, and the position of the gleyed horizons in relation to the podzolised horizons. A further complication to podzol morphology arises from polygenesis, whereby podzolised soils may persist through different environments, especially changes in climate. Among the most common polygenetic podzolised soils are the solodics, which have persisted since a previous, more arid climate. These soils will be discussed in Chapter 12 (relict soils). Fig. 8.1 classifies the main types of podzolised soils on the basis of the six processes of podzolisation, but excludes hydromorphic podzolised soils and solodics.

The variations in the processes and the morphology of the podzolised soils listed in Fig. 8.1 may be accounted for by the five factors of soil formation. The different types of podzolised soils frequently grade into each other, and one type may develop into another as maturity is approached; for example, a grey-brown podzolic may change into a podzol. The names used in Fig. 8.1 are those of Great Soil Groups and represent mature soils. Other soils, such as grey wooded soils or brown forest soils, with less well-developed podzol characteristics, could also have been included, but were omitted for greater simplicity, as it is not intended in this book to discuss every recognised Great Soil Group. The podzolised soils selected in Fig. 8.1 cover the range of processes and morphologies within this soil suborder, though not every soil in the Great Soil Group within the suborder is included. Each of the soils listed in Fig. 8.1 is discussed separately in the following pages. The profile descriptions presented in Figs. 8.2, 8.4, 8.6, 8.7 and 8.9 are of American and Australian podzolised soils, and though they may be considered typical of their Great Soil Groups, there is considerable variation in morphology within each Great Soil Group, and gradation between Great Soil Groups.

PODZOLS

Podzols develop under the optimum conditions for podzolisation, where climate is sufficiently cool to prevent the thorough

Location: North-eastern New England, U.S.A. | **Land-use:** Coniferous forest cleared for farming.

Drainage: well drained. | **Slope:** flat.

Parent Material: glacial till. | **Average annual rainfall:** 42".

Horizon	Depth	Lower Boundary	Colour	Moisture	Texture	Structure	Organic Matter	pH	Special Features
Ol	5"-3½"	gradational, level	brown	moist		leaf structure preserved	horizons of pure organic matter	4.0	loose, freshly fallen leaves
Of	3½"-2"		dark brown						partially decomposed leaf litter
Oh	2"-0"		black			fine granular			greasy, thoroughly decomposed leaf litter
Ah	0"-½"	sharp, irregular	dark brown		loamy sand	apedal single grain	strong root penetration	4.5	horizon discontinuous
E bleicherde	½"-5"		light grey					4.5	
orterde B1fe, h	5"-10"		dark reddish-brown			apedal massive	humus deposition	4.5	⅛"-¼" iron concretions scattered throughout
B2fe	10"-17"		mid yellowish-brown			fine crumb		5.0	
B3fe	17"-23"		light yellowish-brown			fine crumb		5.0	
B/Cfe	23"-28"		light yellow			structure of parent material preserved		5.5	
C	28"+	calcareous glacial till							

Figure 8.2: Podzol, from northeastern New England, U.S.A. (modified from W. H. Lyford, 1946 and 1952).

decomposition of the leaf litter (cf. lateritic soils and krasnozems discussed in Chapter 7), and yet sufficient humification occurs to produce a highly acid humus layer from which large quantities of fulvic acid are released. Fulvic acid is very mobile and assists in the movement of iron and aluminium through the soil profile, resulting in marked horizon differentiation, as seen from Fig. 8.2, in which colour and structure, especially, change abruptly down the profile.

This illuviation of humus and iron often results in cementation of the upper *B* horizon. The term *ortstein* is used for a thin (usually less than 3 inches) firmly cemented horizon (*Bcn*), and *orterde* for a thicker (usually 3 to 10 inches) less firmly cemented horizon (*Bfe, h*). Both ortstein and orterde contain iron nodules or concretions formed by the oxidation and hydration of the iron as it is deposited with the humus, but these are more profilic in ortstein. A silicate clay cemented horizon (*Bt*) which forms sharp boundaries with neighbouring horizons is termed *hardpan*. The terms ortstein and orterde are usually restricted to podzolised soils, but hardpan is widely used in relation to other soils as well.

In podzols along the eastern Australian Coast, the top *B* horizon is frequently very firmly cemented, requiring a hammer for breakage; it is called *coffee rock* (a *Bfe, h* horizon) because of its rich dark brown colour (from the combined deposition of humus and iron). The soil profile described in Fig. 2.2 (Chapter 2) is a podzol containing coffee rock as the *Bfe, h* horizon. This horizon is sharply bounded with the horizons above and below, but the boundaries are very irregular, especially the lower one, which extends deeply along root channels into the *Bfe* horizon which contains illuvial iron but lacks humus.

The podzol from New England, U.S.A., in Fig. 8.2 contains a *Bfe, h* horizon of orterde which also extends irregularly into the B2fe horizon, which, unlike the podzol in Fig. 2.2, contains some illuvial humus as well as iron. In the *Bfe, h* horizons of both these soils, iron has separated out as nodules. Note the yellow colour of the *Bfe* horizons of these two podzols: this colour is caused by the hydration of the iron oxides, as the lower horizons remain wetter for longer periods.

Texture changes are not marked in either of the two podzols, but mechanical analysis usually reveals a slight increase in clay in the *B* horizons, though this may not be apparent in the field texture gradings.

Base removal is thorough in both the podzols, even though the parent material of the U.S. podzol (Fig. 8.2) is calcareous.

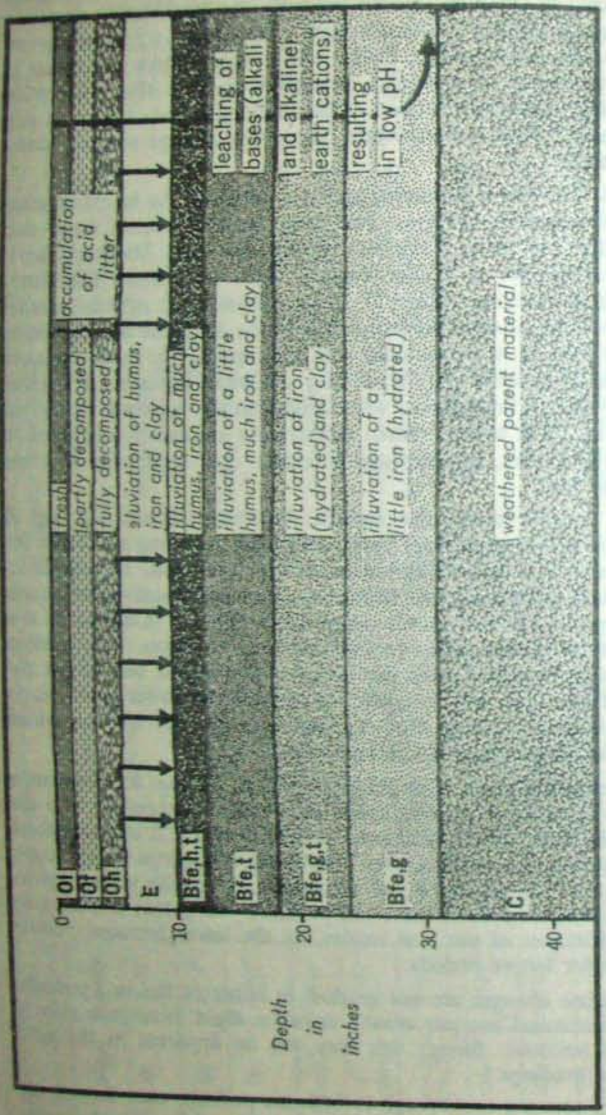

Figure 8.3: The processes of podzolisation leading to the formation of a podzol.

Note how the *pH* gradually increases down the profile in Fig. 8.2 as the calcareous parent material is approached. The low *pH* of podzols makes them fairly unproductive when cropped, but conifers, which have lower nutrient requirements and higher tolerance of acidity than crops, thrive in podzols. After cultivation, the *O* horizons (the main source of nutrients) are frequently lost, exposing the acid bleicherde beneath, which, lacking both iron and humus, is poorly structured and so develops poor tilth. This poor physical condition causes a decrease in yields even in acid-tolerant crops such as potatoes (which are widely grown in podzols) unless the soil is managed carefully. Lime must be applied to podzols before grasses, oats and most other vegetables can be grown successfully.

The podzols in both Figs. 2.2 and 8.2 have undergone all the processes of podzolisation, with perhaps the exception of clay illuviation. These processes are shown diagrammatically in Fig. 8.3 in which a typical podzol profile is sketched.

Podzols are not as well-developed in Australia as in North America or Eurasia, where vast stretches of coniferous (boreal) forest extend across each continent. It is generally too hot in Australia for deep leaf litters to accumulate, the high temperatures causing rapid decomposition of the rather meagre leaf litter supplied by the eucalypts. Compare the thin, poorly developed *Ol* horizon in the Australian podzol shown in Fig. 2.2 with the deep leaf litter divided into three horizons (*Ol*, *Of* and *Oh*) in the American podzol in Fig. 8.2. In cooler, moist areas in Australia, such as western Tasmania and the Monaro region, deeper leaf litters may develop where humus decomposition does not exceed its rate of accumulation. Outside these cooler areas, podzolic soils are more common than podzols, the latter generally being restricted to damp, sheltered locations or to dune sands which podzolise very readily. Australian podzols are generally found in areas with an average annual rainfall of more than 40″; but podzols may still form in cooler parts of the Monaro region where there is a low rainfall, and evaporation is much less. Hallsworth, Costin and Gibbons (1953) describe podzols from both cooler and hotter areas in eastern Australia. One of the most striking differences between the profiles they describe and the Northern Hemisphere podzols is the presence of a definite mineral horizon in the Australian podzols (in Fig. 2.2, a definite *Ah* horizon of sand darkened by humus is present). In the four podzols described by Hallsworth, Costin and Gibbons, the *Ah* horizon varies from two to seven inches in depth, is dark coloured, has a *pH* of 5.0 or less and is similar in texture to the underlying *E* horizon.

In North American podzols, the *O* horizons frequently lie directly on the *E* horizon, with no *A* horizons.

Podzols are found in areas where precipitation exceeds evaporation and it is not hot enough for microbial activity to thoroughly decompose the leaf litter. The main zonal distribution of podzols correlates with Koppen's *Dfc* and *Dwc* climates: these are the coniferous (boreal) forest regions. Podzols are best developed on siliceous rocks such as sandstones or granites. It is doubtful whether true podzols really form on basic rocks such as basalt: the so-called podzols recorded on basalt do not have low *pH* values and usually lack humus and iron illuviation, and so are not even podzolic soils. Limestone probably does not podzolise at all. Deeper podzol profiles are found on siliceous pre-weathered materials like the dune sands shown in Fig. 2.2. Calcareous pre-weathered clayey materials, such as the glacial till in Fig. 8.2, do not form such deep profiles: the podzol described in Fig. 8.2 is comparable in depth to podzols on shale or sandstone in the Sydney region. Free drainage is essential to podzol formation, otherwise a *ground water podzol* will form. This soil contains a gley horizon (bleached, with blue and green colours sometimes appearing, and anaerobic conditions prevailing) somewhere in the profile.

It was pointed out in Chapter 5 that 500 to 1,000 years are necessary for a mature podzol to form zonally. This long period is required because soil formation is active for only part of the year: in winter, the ground is usually frozen.

The podzolic soils develop when the above conditions are not all fulfilled, that is, the conditions are not optimum for podzolisation, and only some of the processes can function.

BROWN PODZOLICS

Brown podzolics are named from the colour of the entire solum. They are also called crypto-podzolics, from the Greek word *kryptos*, meaning hidden. In these soils organic matter dominates the morphology, and some of the features of podzolisation, especially the bleicherde, are hidden from view. In brown podzolics, organic matter decays even more slowly than it does in podzols, so that humus accumulates throughout the profile to the extent that some of the processes of podzolisation are impeded, while others are masked by the soil's dominant dark brown colour. In Fig. 8.1 it can be seen that the only well-developed processes of podzolisation are the leaching of bases and the surface accumulation of an acid organic mat. Some brown podzolics have incipient bleicherdes, usually less than an inch deep, but generally

Location: Windsor, Hartford County, Connecticut, U.S.A.
Drainage: well drained.
Parent Material: stratified glacio-fluvial deposits
Vegetation: deciduous forest with some conifers, shrub under-storey.
Slope: almost flat.
Average Annual Rainfall: 42".

Horizon	Depth	Lower Boundary	Colour	Moisture	Texture	Structure	Organic Matter	pH	Special Features
Ol	2"-1"	gradational, level	brown	moist		leaf structure preserved	horizons of pure organic matter	4.0	loose, freshly fallen leaves
Of	1"-0"		dark brown						partially decomposed leaf litter
Ah	0"-3"		very dark greyish-brown			platy	strongly matted with roots	4.0	
B_1h	3"-12"	gradational, irregular	strong brown	dry	fine sandy loam	weak subangular blocky	high organic matter content	4.5	
B_2h	12"-20"		brown to yellowish-brown			weak crumb		4.5	
C	20"-60"+		yellowish-brown and brown	stratified sand and gravel				5.0	

Figure 8.4: Brown podzolic from Connecticut, U.S.A. (modified from Swanson, Shearing & Bourbeau, 1952).

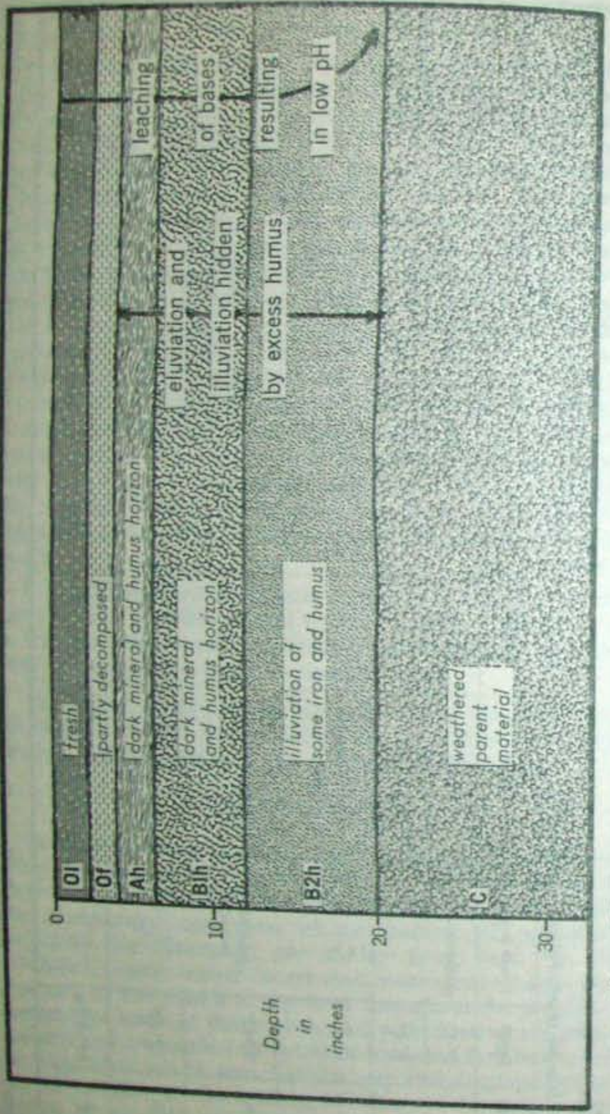

Figure 8.5: The processes of podzolisation leading to the formation of a brown podzolic.

bleicherdes are absent. Humus is often greatest nearest the *O* horizons, decreasing with depth down the profile; so there is no humus illuviation. There is usually little evidence of iron illuviation, the dark brown of the *B* horizons masking any red or yellow colour from iron deposition, but chemical analyses of some brown podzolics reveal a slight increase in iron in the *B* horizons. Clay content is often greatest nearest the *O* horizons, decreasing slightly with depth; or it is fairly uniform down the profile, indicating lack of illuviation. In Fig. 8.4 a brown podzolic from U.S.A. is described and in Fig. 8.5 the processes and morphology of brown podzolics are illustrated.

In Fig. 8.4 note the shallow depth of the profile even though it has developed from pre-weathered material. The horizon boundaries are gradational (cf. the podzols in which there are sharp changes between the horizons), the brownish colour gradually diminishing with depth. Texture is uniform throughout the profile and there is no evidence of orterde or ortstein in the *B* horizons. The *B* horizons of the brown podzolic are not as well-structured as the *B* horizons of the podzol described in Fig. 8.2, except for the massive orterde in the latter. *pH* is fairly uniform throughout the brown podzolic, but increases slightly as the parent material is approached.

In Australia brown podzolics are poorly developed and are restricted to the colder areas, generally more than 4,000 feet above sea level. They have been described in the Monaro region, Western Tasmania and the New England Plateau. As with the podzols, their *O* horizons are generally less well-developed than those in brown podzolics in the Northern Hemisphere.

The conditions necessary for the development of brown podzolics are similar to those for the podzol, excepting that more humus is allowed to accumulate in the profile. This excess of humus may arise either from conditions slightly colder than those in which podzols are found (this applies especially to Australian brown podzolics; in the Monaro, for example, they are found at higher altitudes than the podzols) or from a more substantial leaf litter, such as from deciduous forests. The formation of brown podzolics, as well as podzols, requires excess of precipitation over evaporation, free drainage and a considerable duration of weathering. In the U.S.A., brown podzolics are frequently associated with another Great Soil Group called *brown forest soils*: these soils have even less well-defined podzolic characteristics than have brown podzolics, which may develop from or degrade into brown forest soils.

In brown podzolic soils, conditions are only just suitable for podzolisation. Given an even greater accumulation of humus within the profile, all the processes of podzolisation are impeded: base depletion is hindered and there is no illuviation of constituents. In the Monaro region, at higher altitudes than the areas of the brown podzolics, such conditions prevail and alpine humus soils (see Chapter 9), in which there is no evidence of podzolisation, are found. Thus alpine humus soils, brown podzolics and podzols may be associated in a catena, with the processes of podzolisation increasing with decreasing altitude.

Humus accumulation in podzolised soils may be regarded as a continuum, with brown podzolics at one end and red and yellow podzolics in which humus is lacking at the other, with podzols in the middle:

Maximum humus accumulation	humus decreasing	Minimum humus accumulation
Brown Podzolics	Podzols	Red and Yellow Podzolics

In brown podzolics there is too much humus for humus illuviation to occur and illuviation of both iron and clay are also impeded. At the other extreme, in red and yellow podzolics, there is not enough humus for humus illuviation but clay and iron illuviation are not impeded. In the podzols there is sufficient humus for humus illuviation and this illuviation assists that of iron, but clay illuviation need not be well-developed. Lack of humus, as in the red and yellow and grey-brown podzolics, is associated with clay illuviation.

RED AND YELLOW PODZOLICS

The red and yellow podzolics are found in environments intermediate between those conducive to the formation of podzols and brown podzolics, and those causing ferralisation (accumulation of iron and aluminium throughout the profile). In red and yellow podzolics, iron does not accumulate to such an extent that its illuviation is impeded (cf. lateritic soils and krasnozems); instead, the sesquioxides are strongly illuviated to the *B* horizons. Red and yellow podzolics are named from the colour of their *B* horizons, rather than from the colouration of the whole profile (cf. brown podzolics).

Red podzolics are more commonly zonal soils formed under freely drained conditions. Yellow podzolics form in catenary association with red podzolics, in the poorly drained depressions.

Location: Prospect, 19 miles west of Sydney, New South Wales.

Vegetation: dry sclerophyll forest with sparse grass cover.

Drainage: well drained.

Slope and Aspect: 3°, facing south.

Parent Material and Underlying Rock: shale.

Average Annual Rainfall: 38".

Horizon	Depth	Lower Boundary	Colour	Moisture	Texture	Structure	Organic Matter	pH	Special Features
Ol	¼"-0"	sharp, irregular	brown	dry		leaf structure preserved	horizon of pure organic matter	6.0	horizon discontinuous
A	0"-5"	gradational, level	mid reddish-brown		sandy clay loam	apedal massive	weak root penetration	5.5	
E bleicherde	5"-10"	sharp, level	light reddish-brown					5.0	up to ¼" iron nodules scattered throughout
Bfe, t	10"-20"	gradational, irregular	bright red	moist	medium clay	strong 1" angular blocky		5.0	strong red clay skins
Bfe, g, t	20"-34"		mottled red, yellow and grey			moderate 1" angular blocky		4.5	moderate red and yellow clay skins
C	34"-60"		mottled grey and yellow		stony medium clay	rock structure preserved		5.5	weathering shale
R	60"+	fresh shale							

Figure 8.6: Red Pozolic from Prospect, New South Wales.

They may also form intrazonally on parent materials rich in the more hydrated iron oxides.

From Fig. 8.8 it can be seen that red and yellow podzolics are characterised by base depletion, lack of or minimal development of *O* horizons, a weakly developed *E* horizon (bleicherde), no illuviation of humus, but marked illuviation of both clay and iron. Thus their profiles lack brown colours, aside from perhaps a slight darkening of the upper horizon, and they are strong in red and yellow colours. These features are apparent in the red podzolic described in Fig. 8.6 and the yellow podzolic described in Fig. 8.7, both soils being from the Sydney region.

On many soil maps, red podzolics are regarded as the zonal soils of the central and southern Eastern Australian Coast. The profile described in Fig. 8.6 is typical of these soils. This red podzolic has developed under dry sclerophyll forest, on a freely drained shale slope. Note the sporadic *Ol* horizon which does not decompose into *Of* or *Oh* horizons underneath. The profile is slightly deeper than the podzol and brown podzolic described in Figs. 8.2 and 8.4 respectively, because of the greater intensity of weathering under the warmer conditions. This greater weathering intensity is reflected in the nature of the clay minerals: the kaolin group dominate the red and yellow podzolics, but considerable illite, a 2:1 lattice clay, was found by Swanson, Shearin and Bourbeau (1952) in brown podzolics from Connecticut, U.S.A., such as the one described in Fig. 8.4. In Fig. 8.6, the bleicherde is not strongly bleached and contains scattered iron nodules, the latter being part of the reason for the name, amphi-podzolic, of these soils. *Amphi* means double or doubtful, and is used because some pedologists regard these podzolics as partly the result of weathering in a previous warmer environment when these iron nodules could have formed (double); some pedologists consider these soils to be partly latosolic rather than podzolic (doubtful). Iron nodules in the bleicherde are a feature of both red and yellow podzolics (see also Fig. 8.7), but the bleicherde should be depleted of iron by eluviation, so these nodules could be relict.

The red podzolic in Fig. 8.6 has marked texture-differentiation down the profile, the *A* and *E* horizons being several texture classes lighter than the *B* horizons. The *B* horizons are distinctly redder in colour than the *A* and *E* horizons, indicating iron illuviation, corroborated by the presence of red and yellow clay skins. The increase in both clay and iron in the *B* horizons has led to the development of quite strong peds coated with clay skins, contrasting with the apedal *A* and *E* horizons. The *pH* of the red podzolic described in Fig. 8.6 is higher than the *pH* of the podzol and brown podzolic described in Figs. 8.2 and 8.4, presumably

Location: Wentworthville, 16 miles west of Sydney, N.S.W.				**Land-use:** recreation reserve; introduced grass.				
Drainage: poorly drained.				**Slope:** flat depression.				
Parent Material: alluvium.		**Underlying Rock:** shale		**Average Annual Rainfall:** 40".				
Horizon	**Depth**	**Lower Boundary**	**Colour**	**Moisture**	**Texture**	**Structure**	**Organic Matter**	**pH**	**Special Features**
A	0"-4"	sharp, level	light brown	dry	clay loam	apedal massive	weak root penetration	5.5	
E bleicherde	4"-8"		light grey					5.5	up to ¼" weathered iron nodules scattered throughout
Bfe, t orterde	8"-11"		light yellowish-grey	moist	medium clay	hard, apedal massive		5.5	up to ⅜" iron nodules throughout
Bfe, g, t	11"-26"	gradational, level	mottled grey and yellow			weak, irregular angular blocky		5.0	weak grey and yellow clay skins
C	26"+	stratified alluvium containing horizons of old, buried soils							

Figure 8.7: Yellow podzolic from Wentworthville, New South Wales.

because of the lack of humic acids. However, it is still too low for optimum plant nutrient availability, so red and yellow podzolics have similar cultivation problems to those of podzols, and are particularly responsive to liming.

Throughout the region in which the red podzolic shown in Fig. 8.6 is found, poorly drained areas near creeks are occupied by yellow podzolics, frequently on alluvial parent material. The essential difference between the red and yellow podzolics lies not in the quantity of iron illuviated to the *B* horizon, but in the form in which the iron is deposited in the *B* horizon. Under impeded drainage in yellow podzolics, the iron oxides are strongly hydrated when they are deposited, and so develop yellow rather than red colours. The *Bfe, g, t* horizon of the red podzolic described in Fig. 8.6 also contains yellow iron oxides, indicating that this horizon is less well-drained than the rest of the profile. In the yellow podzolic, the entire profile is poorly drained, so yellow and grey colours feature throughout.

Note the stronger bleaching of the *E* horizon in the yellow podzolic described in Fig. 8.6 compared to the red podzolic described in Fig. 8.7. However, some iron nodules, although weathered, are still present. Except for the colour differences, the yellow podzolic has features similar to the red podzolic—light textured, apedal *A* and *E* horizons overlying heavy textured, pedal *B* horizons.

In Australia, the occurrence of red and yellow podzolics is frequently related to the nature of the parent material. In the Sydney region, much of the Hawkesbury sandstone contains the hydrated iron oxides, so even in well-drained positions, yellow rather than red podzolics tend to form, because of inheritance. Uplifted pallid horizons of lateritic profiles in the Sydney region give rise to yellow rather than red podzolics, whereas the mottled horizons tend to form red podzolics, especially if fragmentary laterite is still present.

Red and yellow podzolics develop as zonal soils on the equator side of podzols, and grey-brown and brown podzolics. They correlate with Koppen's *Cfa* climate and with some areas of *Cfb* and *Csa* climates. In the Northern Hemisphere, they are found in Southeast Asia, around the Mediterranean Sea and in South-eastern U.S.A., that is, in much warmer regions than the podzols. Under these conditions, clay formation is rapid and many pedologists attribute the high clay content of the *B* horizon to clay formation *in situ* as well as to clay illuviation. Thus the red and yellow podzolics contrast with the other podzolised soils in which clay content is much lower. The depth of weathering and the

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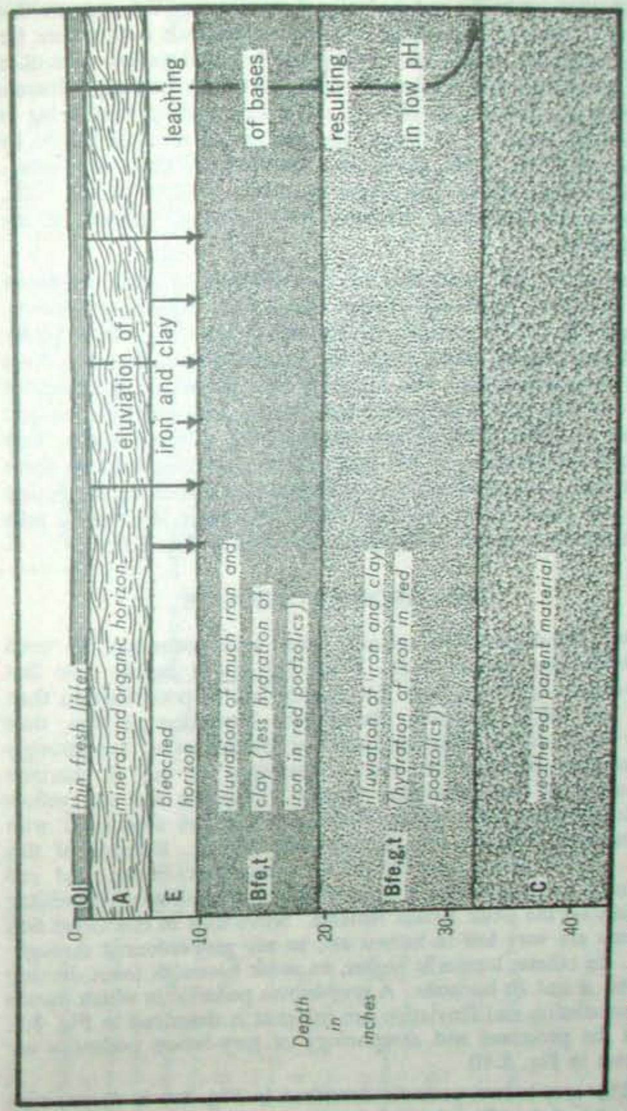

Figure 8.8: The processes of podzolisation leading to the formation of red and yellow podzolic soils.

clayey nature of the *C* horizon (even though it is beyond the depth for illuviation) in the red podzolic shown in Fig. 8.6 support this theory. Chemical weathering of primary minerals is therefore far more important in the morphology of red and yellow podzolics than it is in podzols where clay formation from primary minerals is minimal. Formation of clay minerals by deep-weathering is particularly favoured in the red and yellow podzolics of U.S.A. by Simonson, 1949, who claims that the clay minerals are subsequently destroyed in the upper horizons of the profile; this, coupled with eluviation, accounts for the lighter texture of the *A* horizons.

Red and yellow podzolics are associated more with deciduous and evergreen hardwood forests than with coniferous forests, though in the Northern Hemisphere they do occur under coniferous forests. Their frequent association with hard-leaf trees such as eucalypts, and with Mediterranean woodland, accounts in part for their poorly developed leaf litters. Like podzols, red and yellow podzolics develop more readily from siliceous rocks; very basic rocks in areas of red and yellow podzolics tend to form intrazonal non-podzolised, heavy-textured soils such as the prairie soils and black earths (see Fig. 11.6) on dolerite less than a mile from the red podzolic at Prospect described in Fig. 8.6.

GREY-BROWN PODZOLICS

Grey-brown podzolics are also called lepto-podzolics, the word *lepto* meaning thin or weak. From Fig. 8.1 it can be seen that grey-brown podzolics have one feature less of podzolisation than red and yellow podzolics, namely the illuviation of iron, thus only three of the listed processes of podzolisation are strongly developed—leaching of bases, formation of an *E* horizon (bleicherde), and illuviation of clay. Like the red and yellow podzolics they have poorly developed features associated with humus—*O* horizons and illuviation of humus. Because of this and their minimal iron illuviation, they lack brown and red colours, so their colour is mainly drab brownish-grey, depending mainly on the weak humus features. Some soils in this Great Soil Group are very low in humus and so are grey-coloured throughout. In others, humus is higher, so weak brownish tones develop in the *A* and *Bt* horizons. A grey-brown podzolic in which humus accumulation and illuviation are minimal is described in Fig. 8.9, and the processes and morphology of grey-brown podzolics are shown in Fig. 8.10.

The grey-brown podzolic described in Fig. 8.9 is deeper than the podzol described in Fig. 8.2, but the boundaries between the

Location: Central New England, U.S.A.
Land-use: coniferous forest cleared for farming.
Drainage: well-drained.
Slope: flat.
Parent material: glacial outwash.
Average Annual Rainfall: 38".

Horizon	**Depth**	**Lower Boundary**	**Colour**	**Moisture**	**Texture**	**Structure**	**Organic Matter**	**pH**	**Special Features**
Ol	½"-0"	gradational, level	brown			leaf structure preserved	horizon of pure organic matter	5.0	horizon is discontinuous leaf litter
A	0"-3"	gradational, irregular	greyish-brown	dry	loamy sand	crumb	moderate root penetration	6.0	
E bleicherde	3"-14"		pale grey			apedal single grain		5.0	
Bt	14"-17"		mid yellowish-brown		sandy loam	weak crumb		5.5	horizon discontinuous
Bg,t	17"-32"	sharp, irregular	mottled light yellowish-brown			subangular blocky		6.0	
C	32"+	sandy, calcareous, banded, glacial outwash							

Figure 8.9: Grey-brown podzolic from central New England, U.S.A. (modified from W. H. Lyford, 1946, and P. C. Stobbe, 1952).

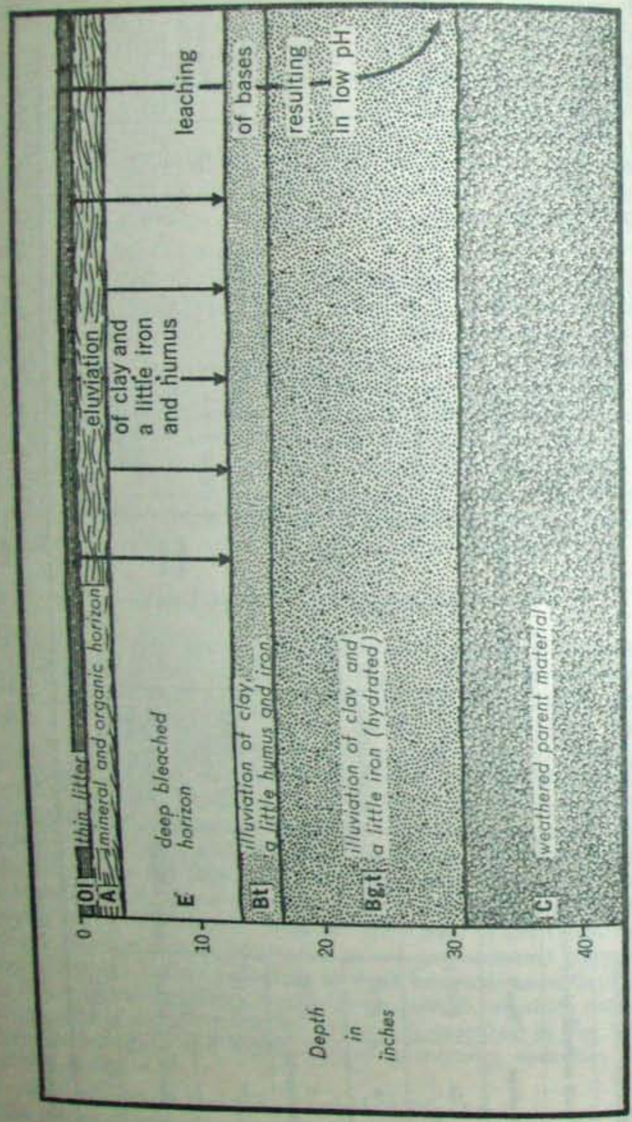

Figure 8.10: The processes of podzolisation leading to the formation of a grey-brown podzolic.

horizons are gradational. The leaf litter is poorly developed and discontinuous, but the *A* and *Bt* horizons are slightly darkened by humus. Texture changes are less marked than in the red and yellow podzolics described in Figs. 8.6 and 8.7; however, the *B* horizons are distinctly heavier textured than the *A* and *E* horizons. As in the red and yellow podzolics, the *B* horizons are better structured than the *A* and *E* horizons and the *pH* is not as low as in the podzols, presumably because of the lack of humic acids from the leaf litter. No iron nodules are recorded in the grey-brown podzolic, but the *B* horizons are yellowish in colour, suggesting that some iron is present. In all, the podzolisation processes are poorly represented in this soil.

Because the features of podzolisation are so poorly developed in grey-brown podzolic soils, it has been suggested by pedologists that these soils are immature brown podzolics or podzols. Stobbe (1952) suggested that the sequence of profile development on calcareous glacial deposits in Eastern Canada is from brown forest soil to grey-brown podzolic to brown podzolic to podzol. He found that brown forest soils occurred in areas which were subjected to the shortest period of weathering; after more prolonged weathering the other soils developed. Stobbe noted that the grey-brown podzolic soils were always found on calcareous glacial deposits, and that non-calcareous glacial deposits formed brown podzolics and podzols.

The zonal belts of grey-brown podzolics and brown forest soils occur between the podzols and red and yellow podzolics in areas with Koppen's *Cfa*, *Cfb*, *Cwa*, *Dfa*, and *Dfb* climates. The main zonal areas of grey-brown podzolics are north-eastern U.S.A., central Europe, and northern China, Manchuria and Japan. In the Southern Hemisphere they are found along the southern tip of Africa and in New Zealand. Grey-brown podzolics are important intrazonal soils in all areas of zonal podzols and red and yellow podzolics. They are found in areas where conditions are unsuitable for the accumulation of deep leaf litters (because of high decomposition rates or meagre supply), especially on iron-poor parent materials.

In Australia, Hallsworth, Costin and Gibbons (1953) consider that grey-brown podzolics develop under conditions of less severe leaching than any of the other podzolised soils, so are usually found in drier, often warmer environments. Drier conditions may account for the minimal mobilisation of sesquioxides. Grey-brown podzolics are found especially along the Eastern Highlands in drier areas and on iron-poor parent materials.

Any of the podzolic soils may be invaded by the water table in any part of the profile, so that the horizons below the water table take on gleyed characteristics. Such soils are termed *gley podzolics* and may develop from brown, grey-brown, red or yellow podzolics.

Podzolisation is the dominating process in cool, moist regions. Considerable variation in soil morphology arises from only small variations in any of the five factors of soil formation. The podzolised soils have the strongest horizon differentiation of all zonal soils, with the exception of lateritic soils. This strong differentiation results from the formation of mobile humus complexes which assist illuviation of other constituents through the profile. Contrast the marked horizon differentiation in the podzolised soils with the lack of differentiation in the soils discussed in the next two chapters: this lack of differentiation is caused by too much humus, or too much clay, either of which impede the process of podzolisation.

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Chapter Nine

Soils Dominated by Organic Matter

CONTENTS

'ORGANIC MATTER' IS THE TERM WHICH DESCRIBES PLANT AND animal remains in the soil. In this chapter, only plant remains are discussed. Organic matter decomposes into a dark amorphous substance called humus, which varies in properties according to the intensity of the decomposition processes, which are called humification. A continuum exists in the humification processes in relation to weathering intensity, as shown here:

weathering intensity and humification processes increase
HOT WET ←———————————— COLD DRY
amount of organic matter in soil decreases

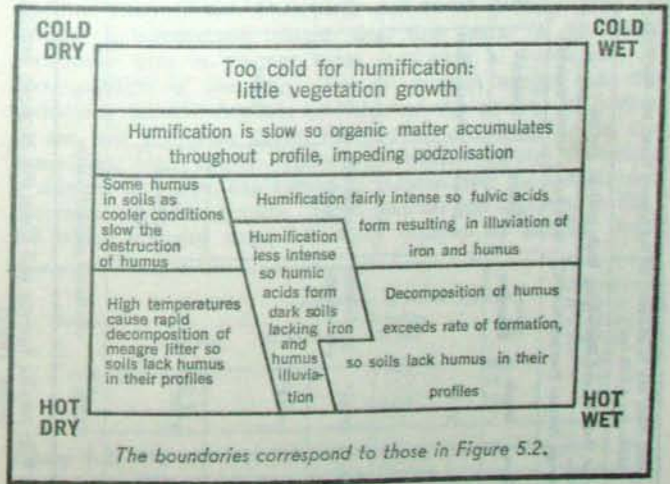
humification is so intense that humus is destroyed more quickly than it forms	humification is fairly intense, so humus rich in fulvic acids is illuviated through the soil	humification is slow, so humus rich in humic acids accumulates in the soil	humification is very slow so little humified organic matter accumulates in soil
e.g., lateritic soils and krasnozems	e.g., podzols	e.g., prairie soils and chernozems	e.g., alpine humus soils

Organic matter thus plays many different roles in soil formation. It is relatively unimportant in the processes of laterisation and ferralisation in the sesquioxide-dominated soils (Chapter 7) as intense tropical weathering destroys the humus more quickly than it forms; hence these soils lack brown and black colours. By way of contrast, in the podzolised soils (Chapter 8) organic matter not only has a visible effect on the soil profile but plays an active role in the podzolisation processes—the fulvic acids assist the mobilisation and deposition of the other constituents, especially iron and aluminium, and so marked illuviation results. Organic matter has quite a different role in chernozems (see Chapter 10)—the stable humic acids impede illuviation and give the soil a black colour which decreases with depth.

In both the podzolised soils and chernozems, organic matter participates in the formal processes of soil formation, namely podzolisation and calcification respectively. However, in brown podzolic soils, where there is much greater accumulation of organic matter than in other podzolised soils, organic matter impedes rather than assists the other processes of podzolisation. The effects of this impediment are visible in the profile morphology: the *E* horizon is lacking, the profile is brown throughout, and there is minimal illuviation of iron, humus and clay. If organic matter accumulates to an even greater extent than in brown podzolic soils, it might be expected that the processes of podzolisation might be hindered completely. This is what happens in the soils discussed in this chapter (soils dominated by organic matter): there is so much organic matter present that the other soil-forming processes, such as illuviation of iron, are unable to function, although precipitation may be adequate for such processes.

The role humus plays in soil formation depends on the factors of soil formation, especially climate, vegetation and relief. Humus conducive to podzolisation is associated with coniferous forests in cool, moist climates, but other forests produce litters which become part of soils with some podzolic characteristics. Since trees feed less heavily on bases than do grasses, forests produce a more acid litter which decomposes into acid, mobile humus[1]. The humus[2] found in chernozems, on the other hand, forms in warmer drier climates supporting grassland vegetation. Grasses, feeding heavily on bases, especially calcium, return them to the surface to form a less acid litter which decomposes into more stable humus. In cold, moist alpine regions, precipitation is ample for forest growth but the low temperature permits only grasslands, or woodland with a well-developed herbaceous under-storey. The high return of bases to the surface is partly offset by the high precipitation, causing strong leaching; however, the low temperatures prevent rapid humus decomposition, so humus[3], in excess of that required for podzolisation, accumulates within the profile. In Australia, such soils are called *alpine humus soils.*

Soils dominated by organic matter also form in environments at high latitudes with climates not unlike those of the high altitude alpine grasslands, although considerably drier. In these areas soils dominated by organic matter are quite common. In Fig. 5.2 (the model of soil zonation in response to climate) a belt of soils, called peat and gley soils, extends from west to east, north of the belt of podzols. In these soils podzolisation is absent and many types of soils dominated by organic matter are found, the types depending mainly on differences in temperature, rainfall and drainage. Alpine humus soils require free drainage and quite high precipitation, and so are less likely to be found in this zonal belt where precipitation is generally less than 30″ per annum. Other types of soils dominated by organic matter tend to develop: if the drainage is impeded and the precipitation is sufficient to cause base removal, *acid moor peats* and *peaty gley soils* form; but if the precipitation is lower or the ground water is alkaline, *fen soils* (alkaline peats) form instead. Thus organic matter-dominated soils may develop in response to both altitude and latitude. The zonal effects of climate (the response to latitude) on the role of organic matter in soil formation are summarised in Fig. 9.1. Note that the boundaries in this zonation model correspond to



The boundaries correspond to those in Figure 5.2.

Figure 9.1: The role of organic matter in soil formation in response to climate.

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[1] The term *mor* is frequently used for the acid litter from softwood forests. This litter is usually sharply demarcated from the mineral horizon underneath (Plice, 1945).

[2] The term *mull* is frequently used for the less acid litter from grasslands or from hardwood forests with a well-developed herbaceous under-storey. This litter is usually well mixed with the underlying mineral horizon (Plice, 1945).

[3] Since this humus is not very acid and merges with the underlying mineral horizon, it would be *mull* rather than *mor*.

Location: Near the headwaters of the Snowy River, Mt. Kosciusko, elevation 6,370 feet.						**Vegetation:** Tall alpine herbfield, dominated by snow grass (Poa caespitosa).			
Parent Material and Underlying Rock: gneissic granite.						**Drainage:** well-drained.			
Average Annual Precipitation: 82", with heavy winter snowfall.						**Slope:** very steep, faces northeast.			
Horizon	**Depth**	**Lower Boundary**	**Colour**	**Moisture**	**Texture**	**Structure**	**Organic Matter, Roots**	**pH**	**Special Features**
Of	3"-0"		brown	slightly moist		finely laminar near surface, becoming apedal near A_1h horizon underneath	litter composed mainly of snow grass, leaves and roots		
A1h	0"-11"	gradational and irregular	brownish-black	moist	sandy loam	fine crumb	humus and grass roots abundant	4.8	earthworms common; some gravel
A2h	11"-19"				gritty sandy loam	crumb		4.8	plentiful decomposing granite stones, quartz gravel and biotite flakes
AC	19"-25"		dark yellowish-brown			apedal single grain	little humus and few grass roots	5.2	plentiful granite stones, quartz gravel and biotite flakes
C	25"+	weathered gneissic granite							

Figure 9.2: Alpine humus soil from the Monaro region, New South Wales (modified from Costin, Hallsworth and Woof, 1952).

those in Fig. 5.1 which may be referred to for the names of the soils within each section.

Like most other Great Soil Groups, the soils dominated by organic matter can also form intrazonally, especially where low relief causes poor drainage, and grass or sedge vegetation rather than forest grows. Intrazonal occurrences of soils dominated by organic matter will be discussed separately with each of the Great Soil Groups in turn.

ALPINE HUMUS SOILS

Until 1952, when Costin, Hallsworth and Woof described alpine humus soils in detail, soils dominated by organic matter had received little attention in Australian pedological literature, perhaps because Australia is too far north to experience the zonal climatic conditions necessary for their formation. However, there are small areas in Australia which have suitable climatic conditions because of their high elevation. The most important of these areas are the Monaro region and Central Plateau of Tasmania, but the highest part of the New England Plateau in New South Wales has some alpine humus soils, though they are not fully developed.

MORPHOLOGY AND SOIL-FORMING PROCESSES

In Fig. 9.2 a typical alpine humus soil profile is described, modified from Costin, Hallsworth and Woof (1952). The *Of* horizon is considerably deeper than that found in Australian podzolised soils in warmer areas, indicating a slower rate of decomposition of the litter. The *Of* horizon merges with the underlying mineral horizon (*A1h*), and the amount of humus in the soil decreases down the profile, corresponding to the diminishing black colour. There is no evidence of clay illuviation in this profile: clay, like humus, is greatest near the surface and decreases down the profile. These features are apparent in Fig. 9.3 which contains some laboratory analyses of samples from

Horizon	% Gravel in Total Soil	Composition of the Non-gravel Fraction			% Organic Matter in Non-gravel Fraction
		% Sand	% Silt	% Clay	
A1h	12.1	57.5	12.4	23.8	6.3
A2h	59.9	62.5	12.1	19.5	5.9
AC	48.2	63.5	18.0	15.6	2.9

Figure 9.3: The size fractions and organic matter content of the mineral horizons of the Alpine humus soil described in Figure 9.2, from the Monaro region, New South Wales (from Costin, Hallsworth and Woof, 1952).

horizons in the profile shown in Fig. 9.2. Notice the high proportion of gravel in the lower horizons; the very high figure for the *A2h* horizon is partly attributable to the presence of a stone line in this horizon.

The high amount of organic matter in the soil confers a stable crumb structure on the *Ah* horizons: ped stability and strength both decrease down the profile as organic matter decreases. The well-developed structure contrasts with that of podzolised soils, whose *A* horizons are usually apedal.

The only feature of podzolisation present in these soils is base depletion, indicated by the low *pH* values in Fig. 9.2. Cation exchange determinations by Costin, Hallsworth and Woof (1952) also corroborate the strong leaching of the profile: they found that the proportion of exchangeable bases within the profile was extremely low, and considered it rather surprising that alpine humus soils are capable of supporting such lush grasses, especially the snow grass (*Poa caespitosa*). The few bases remaining in the soil are more likely to be related to the organic matter than to the mineral colloids, and the strong earthworm activity undoubtedly assists base circulation through the soil.

The dominating soil-forming processes in alpine humus soils, then, are base depletion, the accumulation of a weakly acid mull litter, and the incorporation of humus from this litter into the mineral horizons. The increasing amounts of gravel down the profile indicate that physical weathering, presumably frost-shattering from the frequent freezing and thawing, is perhaps more important than chemical weathering in alpine humus soils (the finer rather than the coarser fractions dominate soils which are chemically weathered). Alpine humus soils even on phyllite, which is fine-grained, unlike gneissic granite (the parent material of the soil shown in Fig. 9.2), contain high proportions of gravel in the lower horizons, indicating that the gravel content is related more to the type of weathering than to the nature of the parent material. The phyllite alpine humus soils contain even more gravel than do the granite alpine humus soils.

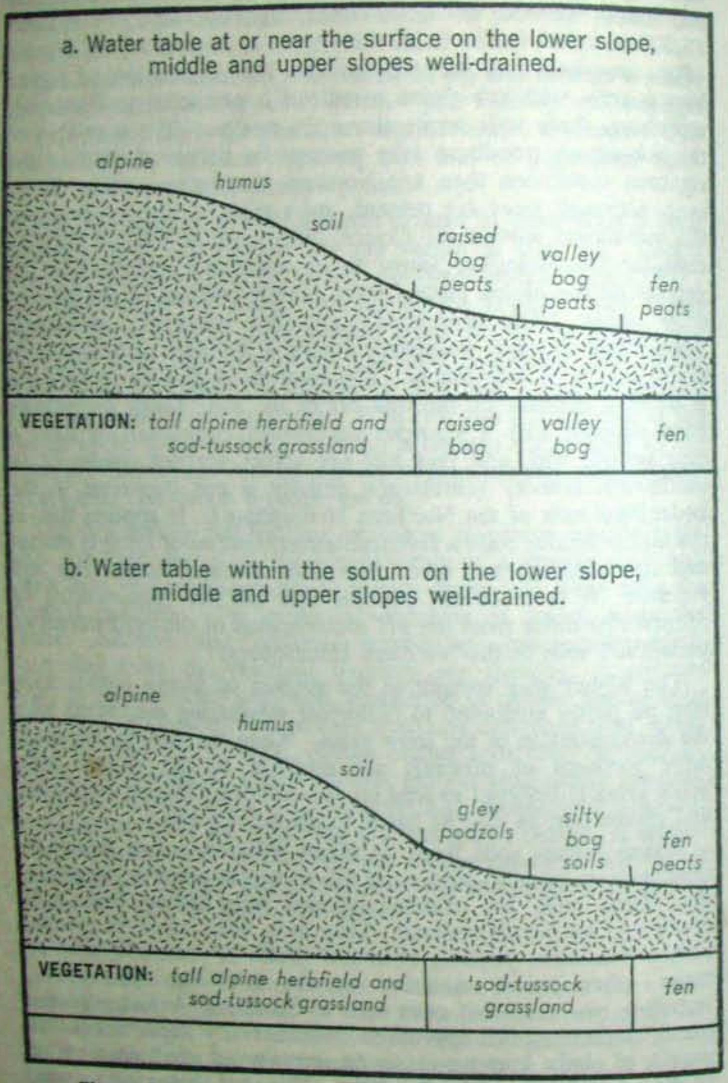
The Factors of Soil Formation

Alpine humus soils are found in distinctive *climates*: the winter snow lies on the ground for at least one month and frequently for six to nine months, but in the summer warmth the ground thaws. In Australia, the average annual precipitation in areas of alpine humus soils varies from 30″ to 90″, most occurring as snow in winter. Severe frosts and wide daily temperature fluctuations occur, but the winter snow protects the soil from these extremes. The heavy snowfalls limit tree-growth, though the quantity of precipitation is ample. Alpine humus soils are associated with tall alpine herbfield (above 6,000 feet altitude), sod-tussock grassland and heath (5,000 to 7,300 feet), and sub-alpine woodland (5,000 to 6,000 feet). In all these types of *vegetation*, snow grass (*Poa caespitosa*) is very important. The association of alpine humus soils with sub-alpine woodland is restricted to Australia; elsewhere, these soils occur above the treeline. The eucalypts of the sub-alpine woodland may perhaps be better adapted to the rigorous conditions than are Northern Hemisphere trees. However, although trees are present, snow grass is still important in the sub-alpine woodland. Costin, Hallsworth and Woof (1952) consider the *biological factor* to be unusually important in the genesis of the alpine humus soils, mainly because of the role of snow grass (*Poa caespitosa*) and the mountain earthworm (*Megascolex*) which frequently attains a diameter of half-an-inch and a length of nine inches. The vigorous earthworm population is unusual considering the low *pH* of the soil. Many pedologists have claimed that a vigorous earthworm population requires a near-neutral soil, and that low *pH* values are not conducive to earthworm activity (earthworm activity is not important in the podzolised soils of the Northern Hemisphere). It appears that in the alpine humus soils a favourable combination of organic matter and soil moisture are more important to earthworms than soil reaction. It is of course possible that the *pH* requirements of *Megascolex* differ from the *pH* requirements of the earthworms of podzolised soils in the Northern Hemisphere.

The higher clay content at the surface of alpine humus soils may be partly attributed to biological weathering associated with the decomposition of the snow grass. When the litter is humified, large amounts of minerals are returned to the surface soil. Snow grass is believed to feed on aluminium hydroxide (gibbsite) and return this, as well as bases, to the surface soil.

Alpine humus soils form on many different *parent materials*, ranging from basic to acidic and fine to coarse grained, suggesting that these soils are climatic climax soils rather than soils dominated by parent material. They have been noted on granite, basalt, limestone, phyllite, slate and gneiss in Australia, and on many other parent materials in the Northern Hemisphere, including pre-weathered ones such as moraine. A factor of the parent material which appears to be universally important in the genesis of alpine humus soils is the accession of small amounts of aeolian dust from arid lands nearby. This dust, being calcareous, is an added factor against the processes of podzolisation.

The nature of the parent material influences the morphology of alpine humus soils far less than it does the podzolised soils where


a. Water table at or near the surface on the lower slope, middle and upper slopes well-drained.

b. Water table within the solum on the lower slope, middle and upper slopes well-drained.

Figure 9.4: Alpine humus soil catenas on gneissic granite in the Kosciusko area, New South Wales.

basic, high clay-forming parent materials strongly resist podzolisation. Alpine humus soils on basalt differ in very few properties from the profiles on granite or gneiss: unexpectedly, they are not higher in either *pH* or clay content than the granite and gneiss profiles, but they do contain more organic matter and silt (from analyses by Costin, Hallsworth and Woof, 1952). The nature of the parent material, then, is unimportant in the genesis of alpine humus soils.

Of far greater importance in the genesis of alpine humus soils is the *relief* factor (especially slope position) and elevation above sea level. Alpine humus soils require free drainage and so are not found in depressions, where peaty and gleyed soils tend to form. At the other extreme, alpine humus soils do not form on very exposed, steep, wind-swept slopes; there, the lithosols tend to form instead.

Alpine humus soils frequently form *catenas* with other soils dominated by organic matter. The alpine humus soils occupy the upper and middle slopes, while peats or gley soils form on the lower slopes. If the water table is at the top of the soil or just below the surface for most of the year, soil formation does not occur, and organic matter accumulates to form peat deposits. On the lowest part of the slope, these peats may be alkaline (fen peats). If, on the other hand, the water table occurs within the solum, soil formation occurs and the horizons below the water table take on gley characteristics (gley podzols); in lower positions on such slopes, soil formation decreases, and there is a development of silty, peaty bog soils or fen peats. These catenary relationships are shown in Fig. 9.4. The vegetation, which changes with the soil down the slope is shown underneath the catenas.

The catenas in Fig. 9.4 occur on slopes without strong micro-climatic temperature inversions, i.e., when there is no appreciable difference in temperature along the slope. Sometimes, however, in the Monaro region, the broad, glaciated valleys are strongly affected by cold-air drainage, while the summits are cold and wind-swept. In such situations the middle slope is the warmest, with less intense frost weathering, which allows podzolisation to operate very weakly. Hence, brown podzolic soils may form on middle slopes, with alpine humus soils on the top and bottom of the slope. The micro-climatic effect is also reflected in the vegetation; on the middle slope this vegetation is wet sclerophyll forest, which produces the acid litter conducive to podzolisation. On the colder top and bottom of the slope, instead of wet sclerophyll forest, there is sub-alpine woodland with patches of sod-tussock grassland, both dominated by snow grass, although there is little change in rainfall over the short distance—temperature is the controlling factor. Such a catena, which is shown in Fig. 9.5, may develop within a vertical distance of as little as 300 feet. The catenas shown in Figs. 9.4 and 9.5 are also known to occur on slate and basalt as well as on gneissic granite.

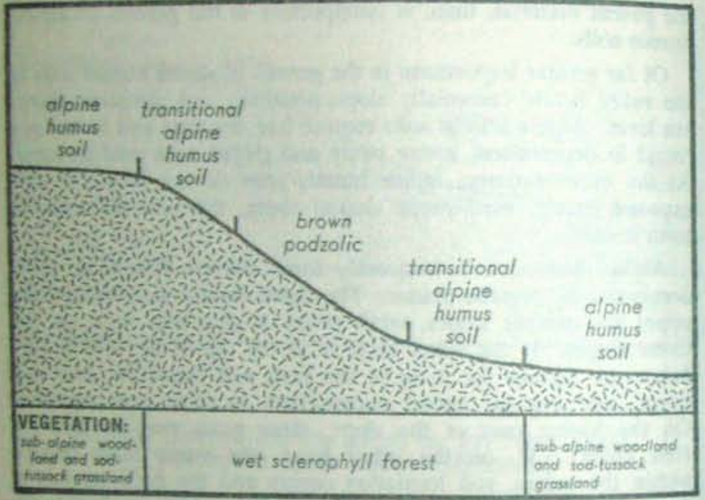


Figure 9.5: Alpine humus soil catena on gneissic granite in the Kosciusko area, New South Wales. The middle slope is the warmest position on this slope.

Elevation above sea level is one of the most important factors in soil genesis in alpine regions. In the Monaro region of New South Wales, this varies from below 3,000 to above 7,000 feet over a horizontal distance of about 20 miles. As rainfall is high in this area, podzolisation occurs wherever the other factors of soil formation are suitable. Basalt in the Monaro region resists podzolisation, to form black earths at low elevations and alpine humus soils at higher elevations. Poorly drained depressions tend to accumulate organic matter, forming peaty, gleyed soils. Very steep slopes form lithosols rather than podzolic soils. Above 4,750 feet generally, or even lower than this where micro-climatic influences prevail, alpine humus soils, not podzolic soils, form, for reasons outlined above. Below 3,000 feet, rainfall diminishes rapidly and soil formation is limited by lack of rainfall, rather than coldness. Fig. 9.6 shows the relationship between elevation, soils, vegetation, temperature and rainfall, in a transect from Jindabyne

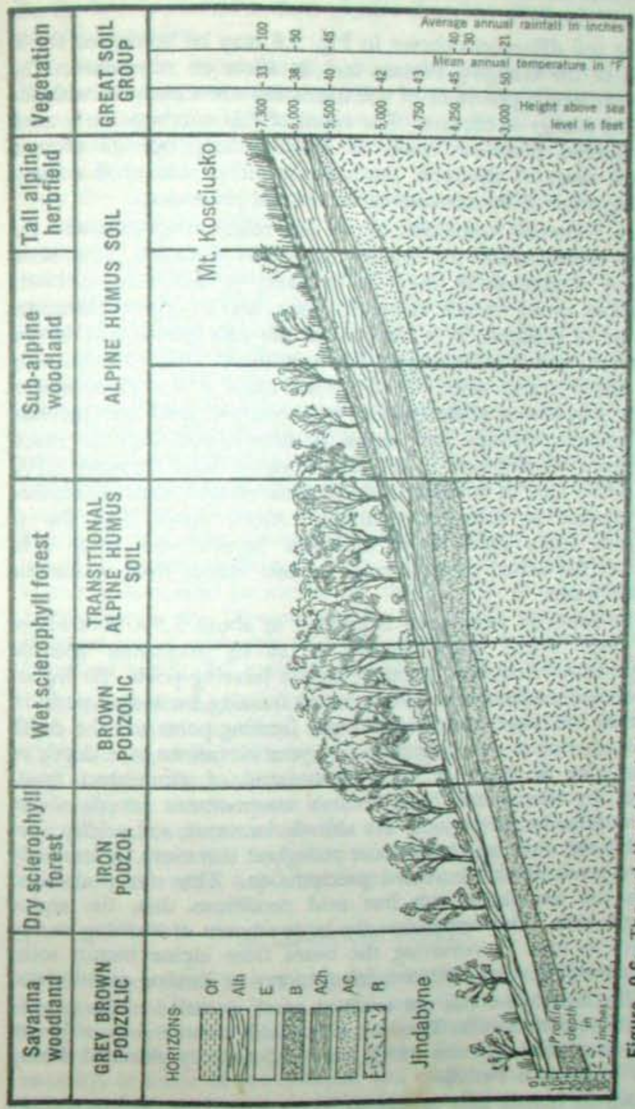


Figure 9.6: The relationship between elevation, soils, vegetation, temperature and rainfall in a transect from Jindabyne to Mt. Kosciusko in the Monaro region of New South Wales.

to Mt. Kosciusko. The underlying rock through this area is mainly gneissic granite.

The soil differences shown in Fig. 9.6 may be accounted for in terms of the changing climate and its effect on vegetation. The different simple processes of soil formation are associated with the marked change in climate. The result of this combination is such that formal processes, such as podzolisation, operate over a narrow range of elevation, the soils on either side of this range lacking some of the features of the formal processes.

The following comments show the relationship between the simple processes and the changes caused by elevation. Illuviation of iron is greatest in the iron podzols; it decreases at lower elevations owing to lack of precipitation, and at higher elevations because of coldness; it is minimal in the grey-brown and brown podzolics; and it is lacking in the transitional alpine humus soils and alpine humus soils. On the other hand, *O* horizons, absent in the grey-brown podzolics and iron podzols, gradually increase in thickness through the region of brown podzolics to reach maximum development in the alpine humus soils between 5,500 and 6,000 feet, as the rate of decomposition of humus diminishes in response to increased coldness. Above 6,000 feet the *O* horizons again diminish in thickness because the bitter cold restricts the actual production of organic matter from which the humus forms.

The depth of weathering is greatest at about 5,500 feet above sea level where frost-shattering is at its maximum because temperatures frequently fluctuate across freezing point. At higher elevations, temperatures remain below freezing for longer periods, so there are fewer fluctuations across freezing point and the depth of weathering is not as great. At lower elevations, the depth of weathering is also not as great because of diminished frost-shattering effects, this time because temperatures remain above freezing for longer periods. As altitude increases, soil acidity rises so that the most acid soils occur at highest elevations, presumably in response to the increased precipitation. Thus the podzolised soils are associated with less acid conditions than the alpine humus soils. This emphasises the large amount of leaching energy which is used in removing the bases from alpine humus soils, presumably leaving little remaining energy to illuviate constituents.

The *time* factor has not received much attention in the genesis of alpine humus soils. They are presumably mature soils, although they are found in areas which were undoubtedly glaciated during the Pleistocene Period.

Alpine humus soils appear to be dominated by climate and the associated vegetation, with relief causing irregularities in the general altitudinal zonation. Parent material and time appear to be far less important in their genesis than the other three factors of soil formation.

In Australia, alpine humus soils are most widespread in the Monaro region; but similar profiles from Central Tasmania have also been described. The alpine humus soils of this region are maximally developed at lower altitudes than those of the Monaro region because of the compensating effect of increased distance from the equator. Maximum development of alpine humus soils is attained above 5,000 feet in the Monaro region, and at least 1,500 feet lower in Tasmania. Of course, alpine humus soils are found at lower elevations than these in both regions, but their properties are less well developed and they tend to grade into brown podzolics. Special conditions of relief (especially aspect, or the direction the slope faces) and micro-climatic influences account for well-developed alpine humus soils below these elevations. They have been recorded as low as 1,500 feet, in the Gippsland region of North-eastern Victoria, under special circumstances. In the New England region of New South Wales, altitude is not sufficient for the alpine humus soils to attain their maximum development as they do in the Monaro region or in Central Tasmania, but quite distinctive profiles are found, particularly on basalt or basaltic colluvium above 4,500 feet. In general, alpine humus soils should be looked for above 2,500 feet in Tasmania, 3,000 feet in the Monaro region, and 4,000 feet in the New England region, although such well-developed profiles as the one described in Fig. 9.2 may not be encountered at these elevations.

In the Northern Hemisphere, soils similar to the alpine humus soils have been described, especially in the Alps and Rockies. They are associated with alpine grasslands rather than woodland, and support some of the better pastures in these areas, used especially for transhumance since they are snow-covered in winter. Although alpine humus soils are not shown on small-scale world soil maps, they are important zonal soils in that they reflect climate, rather than any other factor of soil formation.

ACID MOOR PEATS

Acid moor peats have a wide range of properties, depending on the amount of humus in the profile, and the nature of the underlying mineral material which only partly contributes to the material within the solum. Acid moor peats develop where leaching intensity, in terms of precipitation and evaporation, is quite high, but low temperatures and prolonged saturation of the profile prevent podzolisation. Acid moor peats are frequently found along the edges of alpine lakes where the water table lies within the solum. The profile generally has three horizons: the upper contains fibrous, lighter coloured peat with little mineral matter; the middle contains darker, more humified, greasy peat with some mineral matter; and the lower, which may be below the water table, contains peaty, apedal mineral matter, sometimes with an iron pan. The *pH* throughout the profiles of acid moor peats described in Australia is usually about 5, although coastal organic deposits in Australia and other countries may have much lower *pH* values (Dolman and Buol, 1968, recorded *pH* values as low as 3.5 in peaty soils on the coastal plain of North Carolina). The amount of mineral matter increases and the amount of humus decreases down the profile of acid moor peats. Pedogenesis (soil formation) in these soils is therefore minimal. The main process is the accumulation of organic matter and its slow humification to form an apedal mass (cf. the alpine humus soils in which the lower amount of organic matter produces well-formed peds). The essential environmental difference between acid moor peats and alpine humus soils is poor drainage in the former; this enhances the accumulation of humus. In both soils, low chemical activity occurs.

Acid moor peats are most common in Australia in poorly drained positions associated with alpine humus soils; however, soils in which similar processes operate are found intrazonally in other situations. On the plateaus near Sydney (especially the Woronora to the south) impervious shale lenses within the sandstone often cause the water table to lie very near the surface, and sedge vegetation develops. These areas, though not nearly as high as the alpine regions, are sufficiently cold and wind-swept to keep chemical activity; hence the rate of humification of the sedge vegetation litter is slow, and organic matter accumulates in the soil. Such soils do not usually contain as much organic matter as the acid moor peats of alpine areas, but there is sufficient to prevent podzolisation, even though better-drained soils only a few yards away may have definite podzolic features. The Woronora Plateau near Sublime Point contains these apedal, peaty soils in shallow alluvium at the headwaters of the streams draining to the north. The underlying rock is Hawkesbury sandstone. The surface where these soils develop is fairly flat and is characterised by a lack of trees, and by dense sedge vegetation. The profiles are rarely more than three feet deep and the lighter coloured, fibrous peat of the acid moor peats of alpine areas is generally absent, with only the two lower horizons apparent.

The soil-forming factor, relief, then, is the most important environmental feature of acid moor peats. They may develop on any substrate, for organic matter rather than mineral matter may be considered as the parent material of these soils. The vegetation may be quite variable, the essential feature being the slow rate of decay of the litter rather than the nature of the source of the litter. Sedge and grass vegetation are most commonly associated with these soils. A rather long period of time is necessary for the humus to accumulate, so these soils are found in fairly stable situations.

FEN SOILS (ALKALINE PEATS)

If the ground-water invading the peat profile is alkaline, or if the leaf litter is high in bases, peaty soils will have a neutral or alkaline reaction. Such soils are called fen soils, or neutral or alkaline peats, and are found in Australia on the soft limestones of South Australia. Like the acid moor peats, they are essentially accumulations of humus under impeded drainage, the main morphological difference being base depletion in acid moor peats, and accumulation in fen soils.

Fen soils are found under a wide variety of climatic conditions, including the lowest slopes of alpine regions; however, they are quite common at lower altitudes in drier climates, where they may be reclaimed by draining for cultivation. Their apedal structure poses problems for cultivation as these soils may dry to a very hard mass, so that though the water table is still within the solum, plants may suffer from drought if more than two weeks elapse without rain.

PEATY GLEY SOILS

If the humus accumulation overlies strongly mottled subsoils in which pale colours predominate, the soil is called a peaty gley soil or humic gley soil. In peaty gley soils, the process of gleying is just as important as the process of humus accumulation. Three horizons may be present: the upper peaty horizon with variable amounts of humus; a reddish, yellowish and grey mottled middle horizon through which the water table fluctuates (cf. the mottled horizons of lateritic soils); and a lower, pale horizon dominated by bluish, greenish and grey colours (this horizon being permanently below the water table). Sometimes the water table even rises to within or above the peaty accumulation. Usually the organic matter content of peaty gley soils is lower than that of acid peats or fen soils: this is reflected in the better soil structure and the lower depths of humus penetration. The upper peaty horizon is often crumb or granular in structure, but the peds are not as well developed as those in the alpine humus soils. The mottled horizon and gleyed (lowest) horizon are usually weakly-structured if sandy, and better-structured if clayey. The gleyed horizon, like the pallid horizons of lateritic soils, is usually altered bedrock, with the rock fabric still preserved. Unlike lateritic soils, peaty gley soils are not very deep—seven to eight feet at the most.

The mottled horizon above the gleyed horizons and below the humus accumulation may contain illuvial clay; this, coupled with their base depletion, indicates that these soils have a closer affinity with podzolised soils than the other soils dominated by organic matter, described above. However, humus illuviation is absent, humus being greatest near the surface and decreasing down the profile. With more podzolisation, often the result of polygenesis, these soils may form *moor podzol peats*, a Great Soil Group recognised in Australia by Stephens (1962) but not by Stace, Hubble and others (1968).

This chapter demonstrates the role of organic matter in the various processes of soil formation. Quantity of organic matter is also important in relation to the soil-forming processes: too much organic matter inhibits these processes altogether, as in the acid moor peats and fen soils. If the amount of organic matter is high in relation to the mineral matter, the material ceases to be called a soil and is referred to as *peat* instead[4]. The upper horizons of some acid peat and fen soils would fall into this category, but the alpine humus soils and peaty gley soils contain much less than the required amount of organic matter.

The nature of the products from the decomposition of the organic matter is no less important. Since decomposition is slow and incomplete, the soils dominated by organic matter are higher in humic (the less decomposed fractions of humus) than fulvic acids. This is one of the main reasons why podzolisation is absent from these soils. In this respect, the role of organic matter in alpine humus soils is similar to its role in chernozems, where humic acids also dominate.

All the soils in this chapter, then, have in common large amounts of organic matter incompletely humified; consequently, dark coloured humic acids dominate. Their essential differences relate to the role of the water table in their formation: alpine humus soils are the only freely drained soils; the others all contain the water table somewhere within the solum. If the water table is high in the solum all the year, acid peats or fen soils form, depending on the base status of the ground-water; if the water table fluctuates through the solum and is quite low for part of the year, peaty gley or moor podzol peats form (if polygenesis has occurred).

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[4] In Australia, the limiting figure is about 30% to 50%, varying according to the texture of the soil—higher figures are usually associated with clayey soils.

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Chapter Ten

Soils Dominated by Silicate Clays

CONTENTS

THE SOILS DISCUSSED IN THIS CHAPTER HAVE IN COMMON THE accumulation of silicate clays in all horizons (as distinct from sesquioxide clays which dominated the soils discussed in Chapter 7): that is, there is no marked illuviation of silicate clays (cf. the podzolised soils in Chapter 8), so all the horizons have similar textures. The profiles therefore lack texture differentiation and are clayey rather than sandy. The soils discussed in Chapter 9 also have little texture differentiation down the profile, but for a different reason: the high accumulation of organic matter impedes clay illuviation.

Some of the soils discussed in this chapter may also contain appreciable organic matter, high in humic acids (e.g., prairie soils, chernozems and wiesenbodens); this partly accounts for their lack of texture differentiation. But other soils included here, which lack brown or black colours, contain very little organic matter dominated by humic acids (e.g., shallow red soils and the grey and brown soils of heavy texture). Organic matter cannot therefore account for the lack of texture differentiation in these soils; consequently, another explanation must be sought.

One feature which is common to all the soils grouped here is the nature of their parent materials: they all form on parent materials low in, or lacking, resistant minerals such as quartz, but high in cations such as calcium, iron and magnesium. The parent materials from which these soils are derived, are either inherently clayey (e.g., loess or clayey alluvium) or they are materials which weather readily to form clay minerals (e.g., igneous rocks lacking quartz, and limestone). The soils discussed in this chapter are not found on very siliceous rocks such as quartzose sandstone or granite[1]. These rocks, when found near heavy textured soils, develop quite different soils, depending on the other factors of soil formation. In humid areas, these other rocks are podzolised; in semi-arid areas, they may form red-brown earths.

It was pointed out in Chapter 8 that even though the rainfall may be adequate, certain rocks resist podzolisation. The commonest of these rocks are basic igneous rocks and limestone. Most other rocks form podzolised soils under humid conditions, when forest vegetation produces the acid litter high in mobile fulvic acids, which assist illuviation of constituents. Non-podzolised, uniformly textured soils, then, are found within a very narrow range of parent materials in humid conditions. Under sub-humid conditions, too dry for forest but suitable for perennial grasses, the tendency towards podzolisation is not very strong, as grass litter is basic and high in humic acids. So uniformly textured soils form on a much wider range of parent materials; in fact, they form on all except the most siliceous—diorites and shales. These, which form texturally differentiated soils under humid conditions, readily form uniformly textured soils under sub-humid conditions. Under even drier conditions, with too little rain for a perennial grass cover, illuviation of clay is again possible, as stabilising humic acids are lacking in the soil. Under

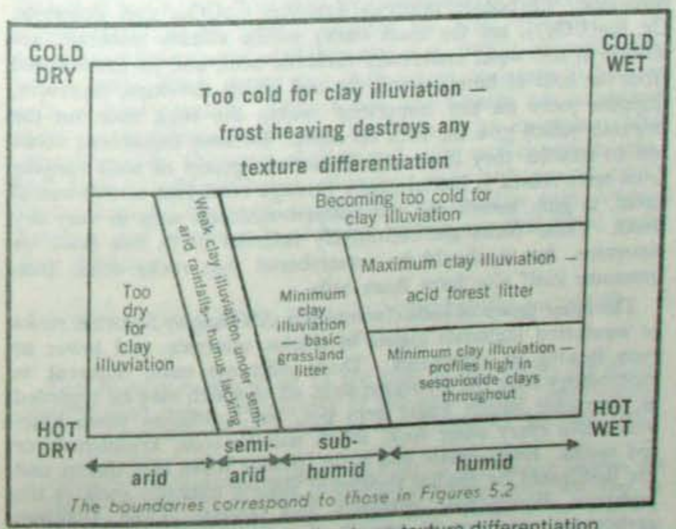

Figure 10.1: The effect of climate on texture differentiation in soils.

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[1] Reported occurrences of these soils on either quartzose sandstone or granite usually reveal mis-classification, or only some of the properties of these soils. Modal profiles, that is, those in which the properties are maximally developed, are not found on these rocks.

these conditions, the range of parent materials which form undifferentiated soils is again restricted to basic rocks and basic alluvium. Since rainfall is low and fulvic acids are lacking, illuviation in semi-arid areas is not very marked. In arid and very cold areas, clay illuviation is minimal, because in arid areas lack of water restricts the movement of the colloids and in very cold areas frost-heaving destroys any texture profile. These ideas are expressed diagrammatically in Fig. 10.1 which shows in model form the relationship between texture differentiation and climate.

From Fig. 10.1 it can be seen that the soils discussed in this chapter attain their maximum development on the widest variety of parent materials in warm, sub-humid areas under grassland vegetation. In this area, their occurrence is zonal; however, these soils are found on either side of their zonal area as intrazonal soils, on a more restricted range of parent materials, such as basalt, limestone and clayey alluvium.

The soils included here may be divided into two groups, depending on the base status of the parent material. The first group of soils forms on rocks very high in bases and low in silicate minerals—highly calcareous rocks (i.e., rocks containing mainly carbonate minerals) such as limestone and dolostone. Carbonate minerals (calcite, $CaCO_3$, and dolomite, $Ca.Mg(CO_3)_2$, are the main ones) unlike silicate minerals, are soluble in soil water containing carbonic acid, and so are leached from the soils of humid areas; the soil which develops, therefore, depends more on the 'impurities' within the rock than on the minerals which give the rock its name. Because calcareous rocks are so soluble, they form a characteristic group of soils ranging from terra rossas in humid areas, through rendzinas in sub-humid areas, to grey, brown and red calcareous desert soils in very dry areas. These rocks are collectively referred to in this book as *limestone*, but it should be remembered that rocks other than limestone itself also form these soils.

The other group of soils discussed in this chapter form on rocks or weathered materials higher in silicate minerals, and lower or even lacking in carbonates. These soils are quite different in morphology from the limestone soils, all of which may be regarded as intrazonal except under very hot, wet conditions when limestone, like every other rock, forms lateritic soils, krasnozems or red earths. Intermediate and basic igneous rocks lack quartz and are dominated by readily weathered silicate minerals such as the feldspars, hornblende, pyroxene and olivine. Lime (calcium carbonate) need not be present, though it does occur in some basic igneous rocks. Although these rocks lack quartz, they still

contain silica as aluminosilicates, but the quantity of silica is much lower than in acid rocks, and cations occupy a large proportion of the rock. These cations, especially calcium from the calcic-plagioclases, may combine with soil carbonates under ineffective leaching to form lime in the soil, even though it may be lacking in the parent material. A wide range of parent materials fall into this category as the only essential properties are lack of, or very little, quartz, and the presence of some basic cations, but not excessive calcium carbonate (cf. limestone which is mainly calcium carbonate). Of the igneous rocks in this category, basalt and dolerite are by far the most common, but diorite, dunite, picrite and other intermediate and basic rocks are also included. Fine-grained, slightly calcareous sedimentary rocks such as calcareous shales, and weathered materials such as loess[2] (which is nearly always calcareous), and alluvium (which may be calcareous) derived chiefly from the rocks mentioned above, are also included. To distinguish this group of parent materials from limestone (the first group of parent materials discussed above), the rather cumbersome term 'basic, heavy textured parent material' is used.

The main zonal soils on basic, heavy textured parent materials, in order of diminishing rainfall, are krasnozems (these soils were discussed in Chapter 7 as they are dominated by sesquioxide clays rather than silicate clays), shallow red soils (resembling the transitional krasnozems of Hallsworth *et al.*, 1952), chocolate soils, prairie soils, black earths and chernozems, and grey and brown soils of heavy texture. Other Great Soil Groups are recognised; these are transitional to the ones listed above; for example, reddish chocolate soils (between shallow red soils and chocolate soils) and brown soils (between chocolate soils and prairie soils). Under impeded drainage, wiesenboden or meadow soils develop, but grey and brown soils of heavy texture are also found under poor drainage.

SOILS ON BASIC, HEAVY TEXTURED PARENT MATERIAL

The diagnostic properties of the soils in this group are rather controversial, hence the naming of the soils is also controversial. This is especially so in regard to the distinction between black

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[2] Loess, which is wind-transported material, is fine-grained because the wind is incapable of moving sand-sized particles long distances in suspension: instead, sand is moved short distances by saltation. Loess consists therefore of silt and clay-sized silicate minerals mixed in varying proportions with calcium carbonate.

earths and chernozems. In this book, preference is given to traditional, overseas usage of the terms rather than to recent Australian usage such as in Stace, Hubble and others (1968). The Great Soil Groups described here differ slightly (in properties) from those with similar names in some other books.

Catenary successions are perhaps better developed in this suborder than in any other group of soils; thus it is quite common to encounter three or four of the Great Soil Groups within half-a-mile along a single slope, so it is unrealistic to consider these soils solely on a zonal basis, separating them from their catenas. It is quite true that there are vast expanses of chernozems in Russia and North America, for in these areas widespread deposition of loess is associated with most of the areas climatically suited for the formation of chernozems in well-drained positions. In Australia this is not so, as much of the sub-humid areas are quite hilly and contain highy siliceous rocks, except for the black soil plains which hardly compete in size with the loess plains of the Northern Hemisphere. In hilly terrain in the Northern Hemisphere, similar catenary relationships to those encountered in Australia are present in these soils, so the catenary concepts given here have universal application, although they are based on Australian examples. Zonal occurrences of these soils in Australia more frequently refer to the dominant soil along a slope, rather than to vast expanses of the soil concerned. These catenary relationships are demonstrated in Fig. 10.2 (a and b) which presents broad generalisations on the zonal occurrence of, and catenary relationships between, these soils.

In Fig. 10.2*a*, a model slope is presented to show the general catenary succession of these soils. Of course, one would be most unlikely to encounter all these soils on a single slope because of the restricted climatic conditions required by certain members of the group. Krasnozems are found only in the wettest areas; if they occur in drier areas, they are relict soils as in the Inverell area where rainfall averages 30″ per annum; grey and brown soils of heavy texture are found only in the driest areas. Even if conditions are suitable for a wide range of Great Soil Groups to develop, some may be missing because of the nature of the slope or the parent material.

In Fig. 10.2b very rough zonal distributions of the main Great Soil Groups are given, based solely on rainfall and assuming that the temperature[3] would correspond to that along a transect across

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[3] In Tasmania, where temperatures are appreciably lower, the soils shown in the zonation in Fig. 10.2 extend into areas of lower rainfall, e.g., prairie soils (soils without lime in the solum, but sometimes with lime in the C horizon) are found in areas receiving less than 25″ average annual rainfall.

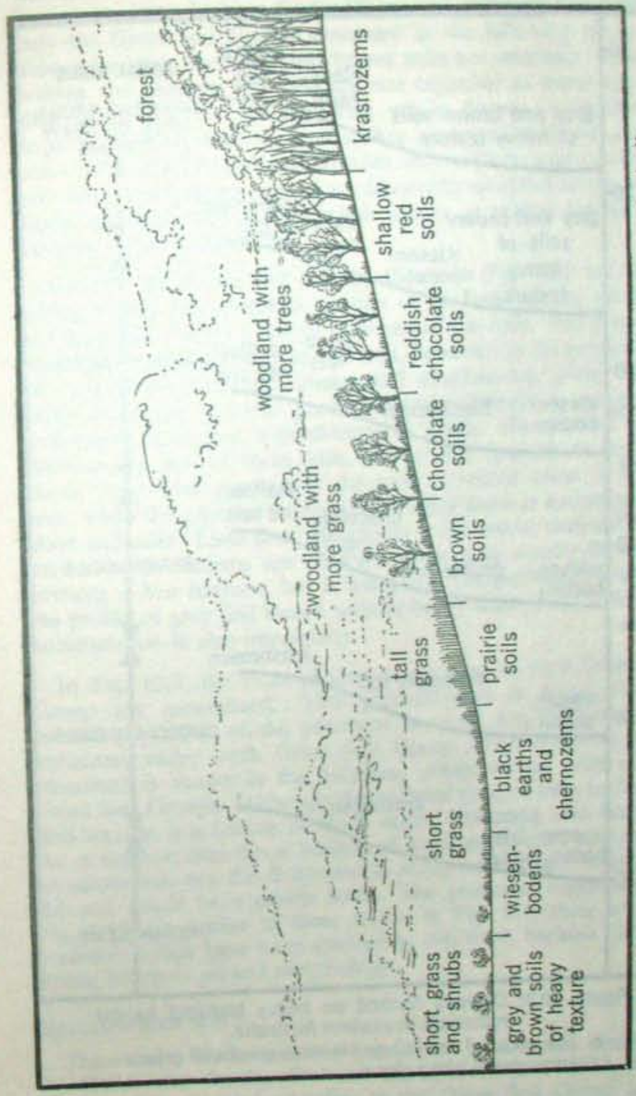


Figure 10.2: Catenas formed on heavy textured parent materials in eastern Australia.

a. Catenary relationship between the great soil groups.

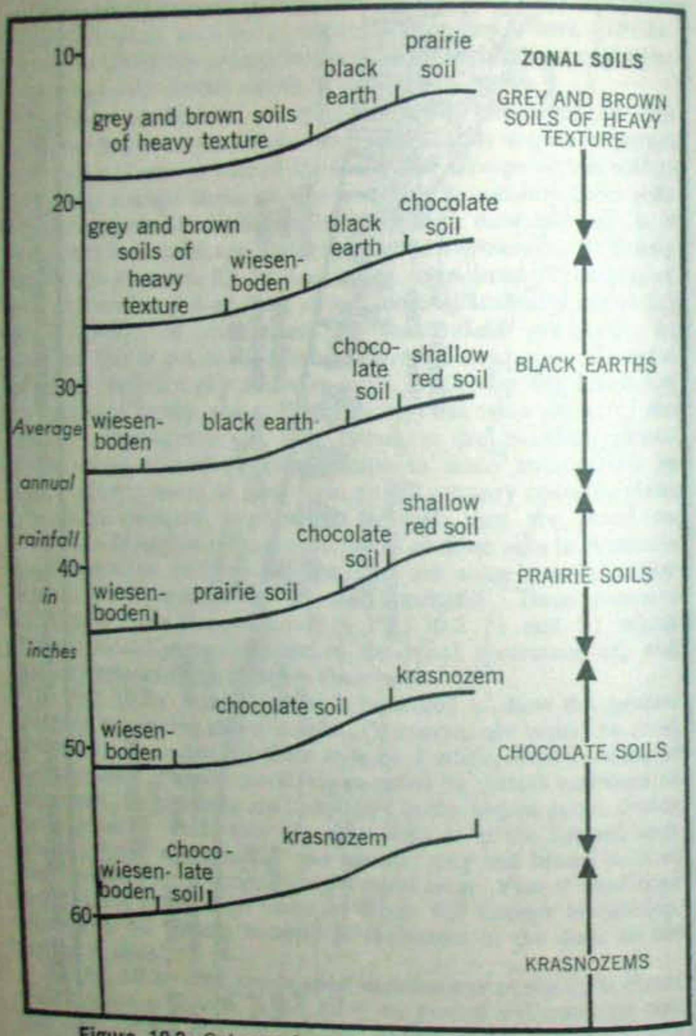


Figure 10.2: Catenas formed on heavy textured parent materials in eastern Australia.
b. The effect of rainfall on the main great soil groups encountered along slopes.

northern New South Wales. Fig. 10.2b also shows some catenas known to occur in New South Wales and southern Queensland: only the Great Soil Groups discussed in the following pages are shown (reddish chocolate and brown soils are omitted). Wiesenbodens and shallow red soils are not regarded as zonal because they commonly do not dominate whole slopes: wiesenbodens require impeded drainage and so are restricted to lower slopes; conversely, shallow red soils require very well-drained conditions and fairly steep slopes and so are frequently confined to the upper slope, although they occupy whole slopes in rugged terrain near Inverell, New South Wales.

Because the Great Soil Groups shown in Fig. 10.2 are found under various environments, certain of the soil-forming processes are restricted to only a few or even one of the soils. The processes which all these soils have in common, however, is the general lack of or minimal illuviation of clay and sesquioxides, giving rise to fairly uniformly textured profiles without changes in redness or yellowness associated with changes in depth. Humus also is not illuviated in any of these soils, and it only features in some of them; the colour pigment in the soils of wetter areas is mainly iron, while the pigment in the soils of drier areas is mainly silicate clays and salts. Lime does not appear in the solum until the black earths or chernozems are reached; there it is usually illuviated, forming a *Bca* horizon; but it frequently accumulates throughout the profile of grey and brown soils of heavy texture (in which salt accumulation is also important).

In Fig. 10.3, the main processes operating in each Great Soil Group are generalised. This generalisation is necessarily an over-simplification of the processes at work, for there are wide variations within each Great Soil Group. The extent of these variations is shown in the following comments on each of the Great Soil Groups. Although the accumulation of lime to form a *Bca* horizon is a feature of black earths, lime may also occur in the *A* horizon; thus the *A* horizon of black earths may or may not be calcareous, but the *B* horizon is always calcareous (otherwise the soil would be a prairie soil). The processes considered in Fig. 10.3 are similar to those shown in Fig. 6.4; these are the processes which have been stressed in this book because of their strong influence on soil morphology.

Shallow Red Soils

These soils, which resemble transitional krasnozems described by Hallsworth, Costin, Gibbons and Robertson (1952), have only recently received attention at the Great Soil Group level.

Great Soil Group	Products of Mineral Weathering				Products of decomposition of organic matter—humus
	Salts of alkali cations, especially Na and K	Salts of alkaline earths, especially lime ($CaCO_3$)	Sesquioxides, especially iron oxides	Silicate clays	
Shallow red soils	removed from the profile	removed from the solum but may be found in C horizon	accumulate throughout in less hydrated form to give red colour	accumulate throughout with weak illuviation to B horizon	humus leached as quickly as it forms so brown and black colours not developed
Chocolate soils				accumulate throughout with no or minimal illuviation to B horizon	chocolate colour arises from iron plus humus which only weakly accumulates
Prairie soils			accumulate throughout in less hydrated form — colour masked by high humic acid content		high humus content in A horizon which is black coloured; humic acids dominate so humus decreases with depth and is not illuviated
Black earths and chernozems	generally removed from the profile, but some may be illuviated to lower B horizons	mainly illuviated to B horizons, but lesser amounts may be found in A horizons			
Wiesenbodens			hydrated in the B horizons to produce yellow colour		
Grey and brown soils of heavy texture	may accumulate throughout; be illuviated to B horizons; or rise to surface and concentrate there	generally accumulate throughout	unimportant		brown soils have higher humus content than grey soils of heavy texture: lack of humus is related to low leaf litter accession from desert vegetation

Figure 10.3: The main processes operating in the most important Great Soil Groups dominated by silicate clays.

Previously all red soils on basic parent material were considered to be deeply weathered, the result of intense, subtropical decomposition. In Australia, red soils on basalt were not regarded as the result of weathering under the present environment unless the average annual rainfall exceeded 55″. Any red soils on basalt found in lower rainfall areas were considered as relict soils from the Tertiary subtropical climate, even though many of the profiles were quite shallow and unlike krasnozems in properties—these differences from true krasnozems were attributed to post-formational changes such as erosion, and the addition of bases. However, it has been noted in Australia and overseas that in well-drained situations on basalt, with woodland rather than tall grass vegetation, the parent rock weathers directly to red soils under rainfalls much lower than 55″, even in warmer areas such as northern New South Wales, where evaporation is high and precipitation is less effective. In steeply sloping areas (slopes of more than 6°) these red soils occupy the entire slope, but in less sloping areas, they give way to chocolate and prairie soils on the lower slopes (see Fig. 10.2).

The shallow red soils have more pronounced eluvial features than the other soils described in Fig. 10.3. There is usually a change in texture between the *A* and *B* horizons because some clay is illuviated, but far less than in some podzolised soils. In this respect the shallow red soils resemble krasnozems which also usually have some texture contrast. Shallow red soils forming the upper members of catenas on basalt in northern New South Wales are shown in Figs. 10.4 and 10.5. From these it can be seen that there is considerable variation in properties in shallow red soils, depending specially on slope and rainfall. The shallow red soil from near Merriwa shown in Fig. 10.4 is much deeper because of the lower slope, but has a high *pH* because of the lower rainfall and lower slope than the soil from near Inverell shown in Fig. 10.5. However, both soils contain both *A* and *B* horizons in their solum, distinguished for reasons of texture, structure and clay skins. The depth of the Inverell soil is more typical of this Great Soil Group than is the Merriwa soil, which does, however, demonstrate that this soil has a wide range of occurrence.

Fig. 10.7 shows the percentages of organic matter (assuming organic matter contains approximately 57% carbon) and clay in soil samples taken from the horizons of the profiles described in Figs. 10.4, 10.5 and 10.6. As Fig. 10.7 shows, the shallow red soils are lower in organic matter than the other soils. Because these soils are associated with woodland rather than grassland, the organic matter decomposes into humus lower in humic acids than the humus in black earths; hence the shallow red soils lack

Location: *Slope down to Bow Creek, 5 miles west of Merriwa, N.S.W.*

Drainage: *Both soils well drained.*

Parent material and Underlying rock: *Basalt.*

Average annual rainfall: *23".*

Vegetation and Land-use: *Shallow red soil has dry sclerophyll woodland; black earth has introduced grasses grazed by sheep.*

Slope:

HORIZON	DEPTH	LOWER BOUNDARY	COLOUR	MOISTURE	TEXTURE	STRUCTURE	ROOTS, ORGANIC MATTER	LIME	pH	SPECIAL FEATURES
SHALLOW RED SOIL										
A	0-12"	gradational, level	strong red	dry	light clay	strong ½" nutty	strong root penetration	nil	7.0	up to 8" basalt stones
B	12-30"	sharp, irregular			medium clay	strong ¾" subangular blocky	weak root penetration		7.5	strong clay skins
C	30-80"	weathering basalt								
R	80"+	fresh basalt								
BLACK EARTH										
A1	0-1"	sharp, irregular	black	dry	light clay	strong ½" crumb	strong root penetration	nodules lying on soil surface, some fine earth in soil	8.5	self-mulching horizon
A2	1"-2'				heavy clay	strong 1" subangular blocky			8.5	tongues of black soil extend into B horizon
Bca	2-5'	gradational, irregular	dark brown and white			moderate 1" subangular blocky	weak root penetration	much fine earth and nodules	8.5	strong clay skins
C	5-9'	weathering basalt								
R	9'+	fresh basalt								

Figure 10.4: Catena on Basalt near Merriwa, N.S.W. Field profile descriptions of shallow red soil and black earth.

Location: *Slope down to the Macintyre River, 4 miles east-southeast of Inverell, N.S.W.*

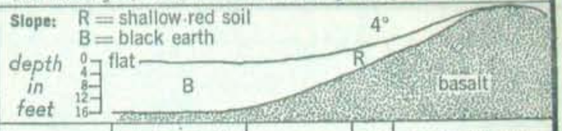
Drainage: *Red soil better drained than black earth.*

Parent material: *Basalt for red soil and basaltic alluvium for black earth.*

Underlying rock: *Basalt.* **Average annual rainfall:** *30".*

Vegetation and Land-use: *Shallow red soil has scattered eucalypts with thick natural grass cover; black earth has sown grasses grazed by cattle.*

Slope:



HORIZON	DEPTH	LOWER BOUNDARY	COLOUR	MOISTURE	TEXTURE	STRUCTURE	ROOTS, ORGANIC MATTER	LIME	pH	SPECIAL FEATURES
SHALLOW RED SOIL										
A	0-3"	merge irregular	mid bright red	dry	light clay	strong ½" nutty	strong root penetration	nil	5.5	up to 6" basaltic stones throughout
B	3-10"				heavy clay	weak irregular polyhedral			5.5	weak clay skins
C	10-60"	weathering basalt								
R	60"+	fresh basalt								
BLACK EARTH										
A1	0-½"	merge irregular	black	dry	light clay	strong crumb	strong root penetration	nil	7.0	scattered basaltic stones and traces of quartz gravel throughout. Strong clay skins from 2" to 10'
A2	½"-3'				heavy clay	strong ¾" blocky			7.0	
Bca	3-6'		black with white			weak ½" subangular blocky	weak root penetration	much fine earth and nodules throughout	8.5	
Bb	6-11'		mottled black and brown	moist		apedal massive	very few roots	scattered nodules from 7'	8.0	
C	11-17'	weathering basalt								
R	17'+	fresh basalt								

Figure 10.5: Catena on Basalt and Basaltic Alluvium near Inverell, N.S.W. Field profile descriptions of shallow red soil and black earth.

Location: *Between blue metal quarry and Prospect Reservoir, Greystanes' property, Prospect, west of Sydney.*

Drainage: *Chocolate and prairie soil well drained, meadow soil poorly drained.*

Parent material: *Chocolate and prairie soil, dolerite; wiesenboden, doleritic alluvium and colluvium.*

Underlying rock: *Dolerite lopolith, intruding shale.*

Vegetation and Land-use: *Introduced grasses grazed by cattle.*

Slope:

Average annual rainfall: 34"

C *chocolate*
P *prairie soil*
W *wiesenboden*

HORIZON	DEPTH	LOWER BOUNDARY	COLOUR	MOISTURE	TEXTURE	STRUCTURE	ROOTS, ORGANIC MATTER	LIME	pH	SPECIAL FEATURES
CHOCOLATE SOIL										
A1	0-½"	sharp, irregular	brown	dry	clay loam	strong ⅛" granular	moderate root penetration	nil	6.0	self-mulching horizon
A2	½"-2'	gradational, irregular	reddish chocolate		light clay	weak ½" polyhedral		nil	5.5	
C R	2-6' 6'+	weathering dolerite fresh dolerite								
PRAIRIE SOIL										
A1h	0-½"	sharp irregular	dark brownish black	dry	clay loam	strong ⅛" granular	strong root penetration	nil	6.5	self-mulching horizon
A2h	½-12"	gradational, irregular			medium clay	strong ¼" polyhedral			6.0	well-developed clay skins
B	12-30"		brown			moderate 1" polyhedral			6.5	
C R	30"-7' 7'+	weathering dolerite containing some lime as weathering calcium carbonate crystals fresh dolerite								
WIESENBODEN (MEADOW SOIL)										
A1h	0-1"	sharp, irregular	black	dry	light clay	strong ⅛" granular	strong root penetration	nil	7.0	self-mulching horizon
A2h	1-10"	gradational, irregular		moist	heavy clay	massive because of high moisture content			7.0	strong clay skins
Bgh	10-30"		black mottled with grey	wet			weak root peneration		8.0	moderate clay skins
Bca	30-54"		dark grey with white				few roots	fine earth and nodules	8.5	
Bb	54"+	series of buried horizons of clayey alluvial soils								

Figure 10.6: Catena on Dolerite at Prospect, West of Sydney, N.S.W. Field profile descriptions of chocolate soil, prairie soil and wiesenboden (meadow soil).

brown and black colours even though they do contain considerable organic matter (the *A* horizon of the Inverell shallow red soil contains nearly 3%, but most of this is poorly humified and consists of fresh roots and leaves rather than humus).

	Horizon	Percentage Organic Matter	Percentage Clay
Shallow Red soil near Merriwa (Fig. 10.4)	A B	1.3 0.8	37.0 52.3
Black Earth near Merriwa (Fig. 10.4)	A1 (mulch) A2 B	2.9 2.8 1.3	43.0 63.5 65.5
Shallow Red Soil near Inverell (Fig. 10.5)	A B	2.7 2.1	50.6 62.8
Black earth near Inverell (Fig. 10.5)	A1 (mulch) A2 Bca Bb	3.8 2.9 1.5 2.7	46.0 71.2 68.2 60.9
Chocolate Soil near Prospect (Fig. 10.6)	A1 (mulch) A2	3.1 2.7	24.8 28.2
Prairie Soil near Prospect (Fig. 10.6)	A1h (mulch) A2h B	5.0 4.2 2.9	32.0 40.0 36.6
Wiesenboden near Prospect (Fig. 10.6)	A1h (mulch) A2h Bgh Bca	7.2 6.0 4.8 3.0	32.3 48.3 59.1 63.4

Figure 10.7: Percentage organic matter and clay in the horizons of soils dominated by silicate clays and described in Figures 10.4, 10.5, and 10.6.

The shallow red soils analysed in Fig. 10.7 are no less clayey than the other soils, and the Inverell profile is actually higher in clay than the prairie profile from Prospect shown in Fig. 10.6. However, the shallow red soils usually do contain less clay, particularly in their *A* horizon, than do their lower catenary associates.

Of the soils considered in Fig. 10.3, the shallow red soils appear to be the most leached because of their *pH*, weak illuviation of clay, lower humus content and lack of lime. This condition is related to their environment, rather than to their parent material, which is similar to that of the other soils. Since shallow red soils are found only on well-drained, generally quite steep slopes, rainfall is very effective as a leaching agent—far more effective than on the lower slopes in the same area. The presence of woodland rather than grassland keeps the return of bases to the soil surface at a minimum (cf. chernozems), further increasing leaching effectiveness. These conditions promote the formation of 1:1 rather than 2:1 lattice clays (e.g., kaolinite rather than montmorillonite), so the shallow red soils are the least heavy and least plastic of the soils being considered in this chapter, though their actual percentage of clay may be quite high. However, although leaching may be effective, the combined effect of slope steepness and lower rainfall prevents the very deep weathering found in krasnozems. The high proportion of basaltic stones found in most shallow red soils (see Figs. 10.4 and 10.5) suggests a constant removal of the upper horizon by colluviation, permitting only a shallow eluvial horizon to develop: this could also account for the absence of self-mulching in shallow red soils.

CHOCOLATE SOILS

Under slightly less effective leaching conditions than those giving rise to the shallow red soils, clay illuviation is minimal, base removal is less effective, and destruction of humus is less complete, resulting in the formation of chocolate soils. These soils derive their name from the brownish-red colour of their solum which is usually poorly differentiated into horizons. In properties, they are really transitional between shallow red soils and prairie soils. A complete continuum of shallow red soils, reddish-chocolate soils, chocolate soils, brown soils and prairie soils may be recognised along a slope, particularly if a trench exposure is available. Like the shallow red soils, chocolate soils are more likely to be encountered on sloping rather than on flat terrain, but they are frequently found on less steep slopes than the shallow red soils, and under vegetation with a higher grass component.

Variations in the morphology of chocolate soils, as with the shallow red soils, can be related to environment. The chocolate soil profile from Prospect shown in Fig. 10.6 is less differentiated into horizons than more typical chocolate soils, perhaps because of the rather steep slope (5°) and the disturbance of the site for grazing, quarrying and water conservation (the doleritic parent material is used for blue metal, and a reservoir is located about 100 yards from the soil site). The natural vegetation, which was presumably woodland, has been replaced by grasses such as paspalum which have changed the properties of the upper horizons, especially the mulch which is not always found in chocolate soils. Like shallow red soils, chocolate soils also have a shallow solum overlying deeply weathered rock and they are less likely to be encountered on pre-weathered parent materials (especially alluvium and loess) other than shallow colluvium. In Australia chocolate soils are restricted to basalt or similar rocks.

The chocolate soil near Prospect (Fig. 10.6) is slightly lower in clay than chocolate soils from northern New South Wales, which may be as clayey as the shallow red soils or black earths (see Fig. 10.7). Although chocolate soils are generally dominated by 1:1 lattice clays, some do contain appreciable montmorillonite, causing them to swell and shrink more than shallow red soils. Unlike krasnozems, which contain only fairly resistant minerals in their sand fractions, chocolate soils and shallow red soils contain large proportions of vulnerable minerals such as olivine, pyroxene, and calcic plagioclase, and lower proportions of iron oxides, indicating that these soils are far less intensely weathered than krasnozems in terms of both their profile depth and constituent minerals.

Chocolate soils are usually well-structured and have a *pH* just below neutral (the Prospect chocolate soil, Fig. 10.6, is more acid than this), making them highly suited for cultivation, since they lack the excess of free iron oxides contained by krasnozems, and to a lesser extent, by shallow red soils. The chocolate soils of northern New South Wales are cropped quite successfully (wheat is grown in the drier western areas, and corn, fodder grasses and other crops in the wetter eastern areas). In the more steeply sloping areas, they are left for rough grazing—like many of the areas with shallow red soils. Both the shallow red soils and chocolate soils are less widespread than the krasnozems, prairie soils, black earths and grey and brown soils of heavy texture, so less attention has been directed towards their fertility than to the fertility of the other soils.

PRAIRIE SOILS

Prairie soils occur intrazonally on basic parent material as do shallow red and chocolate soils, but they are far more important as zonal soils on a wider variety of parent materials, under a native vegetation of tall grassland. Their best-known zonal areas are in North and South America where they lie between latosols and podzolised soils to the east, and chernozems to the west, that is, between the 25″ - 40″ average annual isohyets. These are the tall grass prairies on diverse parent materials, including loess and alluvium.

Morphologically, prairie soils resemble chernozems most closely, the essential difference between them being the removal of lime from the solum of prairie soils, while lime remains in chernozems. Prairie soils may therefore be regarded as transitional between pedalfers and pedocals, although by definition they are pedalfers even if lime is present in the *C* horizons.

The prairie soil from Prospect described in Fig. 10.6 is similar to profiles from North America, although its organic matter content (Fig. 10.7) is less than that of modal Northern Hemisphere profiles. The organic matter is greatest in the mulch, and decreases with depth. Clay content is greatest in the *A2h* horizon, but it decreases in the *B* horizon, indicating lack of illuviation. *pH* is just below neutral, and peds are well-developed in all horizons; thus, both chemically and physically, this soil is very well endowed. The full effects of grassland vegetation are seen in prairie soils: the production of humic acids causes very dark colours in the upper horizons and lack of illuviation of either humus or clay, because the soil exists as stable aggregates. Grass litter provides a heavy return of bases to the surface, and these bases are incorporated into the upper horizons to partly offset leaching and maintain the *pH* close to neutrality. The lack of texture contrast between the *A* and *B* horizons in prairie soils does not imply complete lack of movement of colloids, as most prairie soils have well-developed clay skins which are related to the internal colloid movements which in turn are associated with flocculation and ped development within the horizons.

Prairie soil peds are frequently extremely hard when dry; for example, the prairie soil at Prospect is 100% pedal and its dry peds can hardly be broken between the fingers and thumb. This strong structural development arises from conditions optimum for both flocculation and cementation: the colloid has a high proportion of calcium adsorbed to it (calcium is a strong flocculating agent) and the large amount of stable humus cements the soil floccules.

Chernozems and black earths are well-structured for similar reasons, but chocolate and shallow red soils lack both the flocculating calcium on their colloids, and the stable humus. These deficiences are partly offset by their iron content which assists cementation, but neither of these soils are as well-structured as prairie soils. Krasnozems are well-structured mainly because of their hydrogen domination and very high iron content (much higher than that of the shallow red soils).

Grey and brown soils of heavy texture, on the other hand, are not as well-structured as prairie soils or chernozems, but for different reasons. Sodium (a dispersing agent) tends to have an important effect on their colloids, diminishing the flocculating effect of calcium. Also, even if grey and brown soils of heavy texture were flocculated, there would be very little to cause cementation as both iron and organic matter are lacking.

Although prairie soils lack red colours, they have been found to be quite high in iron, often as high as chocolate or shallow red soils. The lack of red or yellow pigments arises from the domination of soil colour by the humic acid black pigment which masks any other soil colours. Removal in the laboratory of humus from prairie soils usually reveals reddish tones in the treated sample. Chemical analyses of prairie soils do not reveal iron illuviation; this is not surprising, for iron is frequently co-illuviated with humus.

Prairie soils are not usually very deep (the one from Prospect described in Fig. 10.6 could be considered modal); this condition reflects their moderate weathering environments. Within their shallow solum, however, there is very strong pedological organisation: two distinct horizons are usually present, based on colour and structure changes. Naming these horizons *A* and *B* is rather controversial, especially if these letters have an eluvial and illuvial connotation, so some pedologists prefer to designate them as *Ah* and *A*—either system is acceptable. These sola horizons differ markedly from the underlying *C* horizon, even if it is pre-weathered: the prairie soil-forming processes completely destroy any features arising from sedimentation.

Prairie soils are best developed in areas of free drainage (wiesenbodens form in poorly drained areas of prairie soils), which have an average annual rainfall of about 40″, a hot summer which allows the profile to dry, a high clay-forming parent material, and tall grassland vegetation. However, less fully developed profiles are very common where all these conditions do not occur. On headlands along the south-eastern Australian coast, soils similar in morphology to prairie soils develop under special conditions, resembling those mentioned, but with a rainfall which is often considerably higher. These soils are called *black headland soils* (Parberry, 1947) and reflect the interaction of sea and vegetation. They are best developed on south-facing promontories which receive the full force of the south-easterly winds and regular accessions of salty spray which renew soil bases. They are associated with a low shrub (*Westringia rosemariniformis*) and grassland, and also with heavy textured parent materials well supplied with bases. They are therefore best developed on basalt. Their organic matter content is very high: Parberry found more than 17% organic matter in the basaltic black headland soils and more than 10% in soils on other rocks such as tuff. The black headlands soils are seldom more than two feet deep and their *pH* is close to 5.5 in spite of the frequent base accessions from the ocean and the basic parent materials: the low *pH* is undoubtedly the result of the heavy rainfalls (most sites receive 50″ or more per annum). Although the rainfall is high, the soils are greyish-black rather than red, owing to the prolific leaf litter which slowly decomposes into humic rather than fulvic acids. Morphologically, then, the black headlands soils are more like prairie soils than any other Great Soil Group. Along the coast, they are intrazonal, requiring very special conditions.

Excluding the black headlands soils, prairie soils in Australia are commonly lower in organic matter than their counterparts overseas, probably because Australia lacks the tall prairie grass vegetation so conducive to the accumulation of organic matter. Because of the higher temperatures, prairie soils tend to be found in wetter locations in Australia. One of the main areas of prairie soils in Australia is along the south coast of New South Wales from Bega to just south of Nowra where rainfall generally exceeds 35″ per year: these soils rarely contain more than 6% organic matter in their upper horizon. Prairie soils are also found in basaltic catenas in northern New South Wales, southern Queensland and in Tasmania.

Prairie soils are extremely productive: not only do they have all the desirable features of black earths and chernozems, but in addition they are found in wetter regions, so water stress is not such a critical factor in their cultivation. They are particularly suitable for grains and fodder grasses and in New South Wales provide excellent dairying areas, although their distance from the Sydney market poses problems.

Black Earths and Chernozems

The names 'black earth' and 'chernozem' are used differently among pedologists. The FAO/UNESCO (1968) world soil mapping authority prefers the term 'chernozem' to 'black earth' and recognises three varieties: halpic chernozems, clayey soils with illuviated lime in the *B* horizon; luvic chernozems, clayey soils with some clay illuviated to the *B* horizon in which lime accumulates; and calcic chernozems, clayey soils which are calcareous throughout (all three varieties have a dark *A* horizon rich in humus). Stephens (1949) in his paper on the comparative morphology of world Great Soil Groups suggested that 'black earths' be used for the soils lower in organic matter and 'chernozems' for soils higher in organic matter in their surface horizon; he suggested that 4% organic matter be taken as the boundary. Stace, Hubble and others (1968) use 'chernozems' for soils lower in clay content, with clay illuviation, and 'black earths' for soils with more than 50% clay in their solum. In introductory studies on soils, the terms are used synonymously (e.g., in Strahler's *Physical Geography*, 1960). In this book, Stephens' (1949) approach is adopted, that is, 'black earths' refers to soils containing less than 4% organic matter in the upper horizon and

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Bunting (1967) considers that the modal chernozem has a dark surface horizon two feet deep in the U.S.A., three to five feet deep in Russia, and that in both countries it contains from 5% to 20% organic matter dominated by humic acids. Within the dark horizon, Bunting distinguishes the upper part, with *pH* usually about 6.5 and a granular structure, from the lower part, with *pH* usually 7.5, a prismatic structure and less organic matter. The *B* horizon is lighter in colour (it has less organic matter) and lime-rich. The *C* horizon which is also lime-rich, contains less clay.

Black earth profiles in Australia are more akin to the U.S. modal chernozem in depth, although on alluvium their *A* horizons may be up to five feet deep. The prismatic structure of the Northern Hemisphere chernozems is usually lacking in Australian black earths which are generally blocky-structured. Two black earth profiles are described in Figs. 10.4 and 10.5, from Merriwa and Inverell respectively, the latter receiving more rain than the former. Both soils are self-mulching, a property which is most prized, for these soils tend to crack rather badly in dry seasons. Both soils are heavy clays (aside from the mulches, the horizons all contain more than 60% clay: see Fig. 10.7) with a darker horizon containing well-developed blocky peds overlying a lighter horizon with poorer structure and less organic matter (the *A* horizons contain more than 2% organic matter and the *B* horizons less than 2%: see Fig. 10.7). The Inverell black earth (Fig. 10.5) contains lime only in the *B* horizon, and the *A* horizon has a neutral reaction; but the Merriwa black earth (Fig. 10.4) contains lime in both the *A* and *B* horizons, though the *B* horizon is very much richer in lime. Illuviation of calcium to form deposits of free lime in the *B* horizon is therefore common to both soils, but in the Merriwa profile, where rainfall is less, this illuviation is less thorough and so lime accumulates to a lesser extent in the *A* horizon as well. The high amount of lime in the Merriwa soils could also be related to the parent material: the soil has developed from basalt traversed by veins of calcite (calcium carbonate) which remain in position, although they are partly weathered, in the *C* horizon. The parent material of the Inverell black earth is apparently alluvial, indicated by the

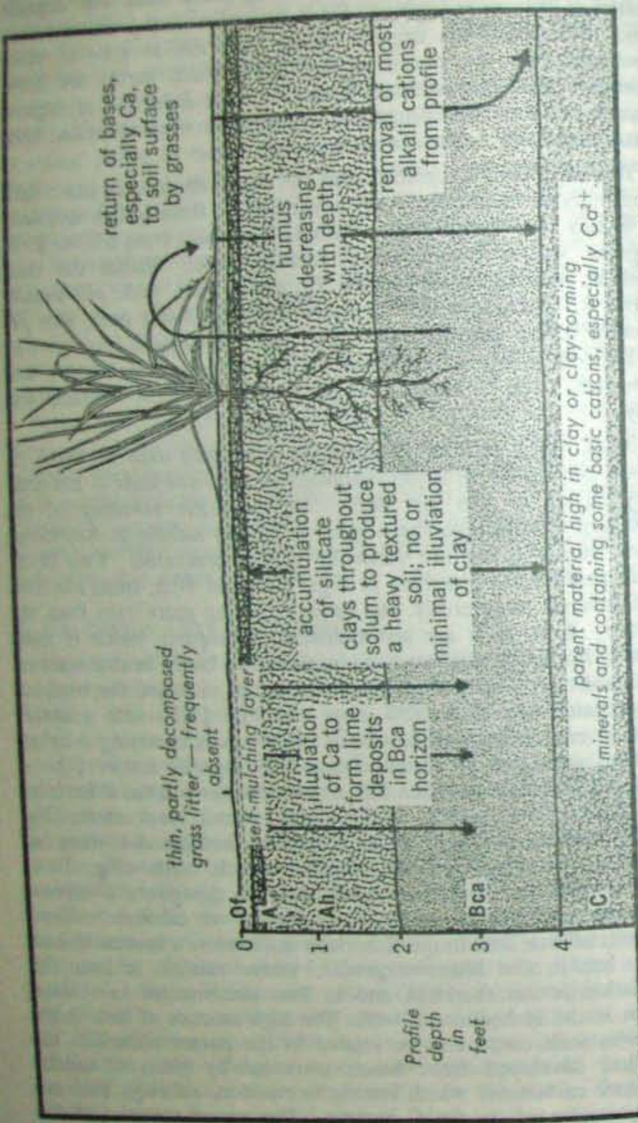

Figure 10.8: The soil-forming processes in chernozems and black earths.

presence of non-basaltic stones in the soil and by non-basaltic minerals seen within in the sand fractions when viewed under the microscope. The Inverell black earth overlies a buried soil which contains more organic matter than the *B* horizon of the black earth: this soil overlies weathering basalt which does not react to hydrochloric acid and lacks calcite veins. The Inverell black earth corresponds fairly well to the halpic chernozem of the FAO/UNESCO system, and the Merriwa black earth to the calcic chernozem.

The processes operating in chernozems and black earths are similar though the quantities of materials accumulating may differ. These processes, which are shown diagrammatically in Fig. 10.8, are intimately related to the grassland vegetation which plays an important role in the formation of humic acids and the return of calcium to the soil surface, leaving the colloids calcium-dominated and the soil stable and flocculated. Humus illuviation is not a feature of chernozems or black earths, for humus is always most abundant close to the surface and it diminishes with depth; but in some chernozems and black earths, weak clay illuviation may occur. More commonly, however, clay also is most abundant near the surface (excluding the mulch) and diminishes with depth. Alkali cations (especially sodium and potassium) are generally removed from both chernozems and black earths, though some may remain on the colloids in the lower *B* horizons. Free alkali salts rarely accumulate in either chernozems or black earths.

The weathering environments of chernozems and black earths are most suitable for the development of 2:1 lattice clays, especially montmorillonite. These clays have the highest shrinking, swelling and cation exchange capacities of all clays and so they further enhance the clayey properties of these soils. Chernozems with as little as 30% clay still feel heavy and plastic, and shrink and swell markedly, and those with high proportions of clay, like the ones discussed here, are extremely difficult to manage (as little as 25 points of rain makes roads on them impassable); they clod badly when they are ploughed wet and are too hard to plough when very dry. However, their high montmorillonite content and frequent self-mulching nature gives them high water-retention ability. Their high humus content, close-to-neutral *pH* in the *A* horizon, high base status, and strong structure make them very productive though they may be difficult to manage immediately after rain. The black soil plains of northern New South Wales (for example, near Moree and Narrabri) and the Darling Downs of Queensland contain some of the most productive cropland in Australia. Compared with other soils, very little fertiliser is added to black earths to maintain yields.

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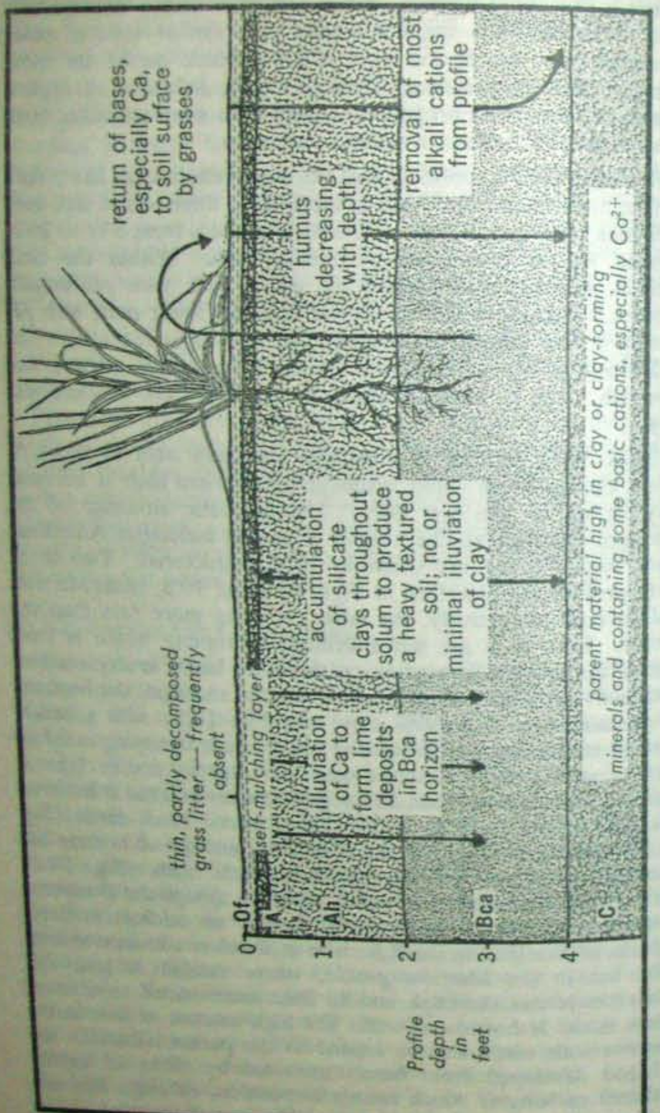

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Recently, black earths have been used successfully for cotton growing near Wee Waa in New South Wales. Some of the best wheatlands are found on the western slopes in rolling black soil country near Inverell, New South Wales. In the same areas, many farms on granite have been abandoned because of poor soils, while the grain farms on the black soil have maintained high yields even though some farmers have not applied any fertiliser.

Black earths are quite unsuitable for orcharding: in fact, trees of any kind do not grow well on them. The treeless nature of the black soil plains of northern New South Wales has not yet been adequately explained though many theories have been proposed. Many of the plains are extremely flat, except for gilgai micro-relief, and contain not a single tree for many miles.

Grazing on black earths poses many problems, because pests which could not survive under poorer conditions thrive in the productive black earths: soil fauna and weeds grow bigger and better in black soils than they do in other soils.

Grain farming, with a fallow to conserve moisture and assist in pest eradication, and also cotton growing seem to be the most successful activities on the black earths and chernozems of most countries. The famous wheatlands of U.S.A. and Russia occupy vast areas of chernozems and prairie soils; the best U.S. cotton lands are on chernozems in Texas; the main cotton areas of India lie on the regur, or black cotton soils developed on the volcanic rocks of the Deccan traps.

The *regur* of India are morphologically similar to Australia's black earths but contain even less organic matter, perhaps the result of both higher temperatures and prolonged cultivation. The regur are found on the Deccan Plateau between the 17.5″ and 53″ average annual isohyets, with more than 75% of the rain falling in the summer monsoon (Hosking, 1935). Tropical black earths like the Indian regur are also found in Africa under savanna grassland, with strong seasonal rainfall concentration. They have been used successfully for cotton in Africa also, for example, in the Gezira (Sudan) and in Nigeria, where irrigation compensates for low and unreliable rainfalls. Black earths are common in all drier inter-tropical areas, especially on lower slopes where they may form catenas with krasnozems. They are by no means restricted to the mid-latitude grasslands, as some textbooks imply.

Black earths and chernozems are therefore found under a wide variety of climatic conditions. In the tropics, they are found in much higher rainfall areas, which in temperate regions would give rise to prairie or chocolate soils. In temperate regions, they are most common between the 20″ and 40″ average annual isohyets, but they also occur on lower slopes in wetter areas. In most instances, they are associated with some type of grassland vegetation: savanna grassland within the tropics, and temperate short grass elsewhere. Seasonal concentration of precipitation causes the grass to grow rapidly for several months, then to die and produce a heavy litter in a very short time. Until the next heavy rains, when the grass is renewed, the main process is the humification of this litter and its incorporation into the soil profile. Grain growing on these soils causes similar processes to occur, since the straw is often left to be ploughed in before the fallow. The procedure with cotton differs: so there is a heavier drain on the soil's natural humus reserves, and they may even become depleted—as is thought to have occurred in the regur (Hosking, 1935).

Because of the strong influence of vegetation on the morphology of chernozems and black earths, parent material is not quite so limiting to their occurrence as it is to chocolate and shallow red soils. However, very acid rocks do not form typical profiles, though they may have some features which resemble chernozems or black earths. Optimum development of chernozems and black occurs on basic igneous rocks, such as basalt, on glacial loess and on basic alluvium. Glacial loess, because of its wind-transported origin, is inherently fine-grained and is also calcareous, so semi-arid to sub-humid areas which are loess-blanketed provide hundreds of miles of virtually uninterrupted chernozems, except in depressed areas, where wiesenbodens are found. It is therefore not surprising that the chernozems of the Russian Steppe were the first to be examined. Black earths are less widespread in Australia: the main areas are either the western slopes in which black earths form only part of catenary successions, or the black soil plains which, though more extensive than the slope black earths, in no way compare with the Russian Steppe, or Central U.S.A., or South-Central Canada.

WIESENBODENS

Wiesenbodens or black meadow soils are intrazonal soils produced by drainage impediment in heavy textured materials. Their profile resembles that of chernozems and black earths, with the exception of their sub-soil horizons which are distinctly mottled with drab greys and yellows, presumably the result of the fluctuating water table.

The properties of wiesenbodens vary according to the levels of the fluctuating water (some wiesenbodens are mottled almost to their surface, whereas others are only mottled at depth) and to the nature of the soils with which they are associated. When prairie soils occur upslope, the wiesenbodens are not likely to contain as much lime as they do when black earths and chernozems occur upslope. They are also associated with podzolic soils if the lower slope parent material is high enough in clay to prevent podzolisation: a common situation is podzolised soils on the hillslopes on siliceous rock and wiesenbodens in the valleys on basaltic alluvium. In this case, they are far less likely to contain lime, for the drainage water from the podzolics would not be so calcareous. The reverse may occur where the ground-water is calcareous: the invading water may provide an even more important source of lime than the parent material of the wiesenboden.

The wiesenboden profile from Prospect described in Fig. 10.6 is associated with a prairie soil immediately upslope, so its lime content is not particularly high. Lime has been illuviated to about 30 inches, accumulating as both nodules and fine earth. There is a slight increase in clay content down the wiesenboden profile (Fig. 10.7) but humus is greatest at the surface and decreases with depth. This soil profile seldom dries out, especially in the lower horizons; estimation of structure is difficult because the soil, containing as it does large proportions of montmorillonite clay, is usually massive. The profile overlies a series of buried soils which extend down to at least 15 feet.

Wiesenbodens are essentially hydromorphic black earths, chernozems or prairie soils. They are not shown on world soil maps because of their intrazonal character and restricted occurrence.

GREY AND BROWN SOILS OF HEAVY TEXTURE

The grey and brown soils of heavy texture[4] are zonal soils, with strong limitations with respect to parent material; more limited, indeed, than chernozems or black earths. They are found in dry or seasonally dry areas on heavy textured alluvium, and so are associated with low relief and periodic flooding. They therefore have both zonal and intrazonal characteristics. Their ubiquity in Australia certainly warrants their consideration as zonal soils, but their hydromorphic features give them intrazonal properties like the wiesenbodens.

In Australia, the grey and brown soils of heavy texture are associated with the large inland river systems, especially the Murrumbidgee-Darling system and the rivers draining towards Lake Eyre. These streams have wide alluvial plains which are, or were (prior to irrigation) periodically inundated as river-discharge is, or was, extremely erratic compared with that of streams flowing to the east coast. Some of the alluvium on which these soils have formed was deposited during the Pleistocene. It has a fairly high clay component, since the greater part of its source area contains fine-grained sedimentary rocks (such as shales) and igneous rocks. The other major areas of these soils in Australia are the Barkly Tableland, situated chiefly in north-eastern Northern Territory and partly in northwestern Queensland, and the alluvium of the streams draining into the Gulf of Carpentaria.

In the northern areas of grey and brown soils of heavy texture, the rainfall is very much higher than in the southern areas: almost all the rain falls during the summer monsoons, when the soils, at first fairly impermeable, are flooded, sometimes to a depth of three feet. These are low-lying areas which provide little opportunity for the water to flow away. During winter, these soils dry out, the vegetation dies, and the salts re-distribute themselves throughout the profile. In the southern areas, these soils are most common in areas which receive less than 20″ of rain per annum; the rainfall, though not seasonally concentrated, is spasmodic. Though northern Australia has a much higher total rainfall, because the rain comes mainly in the summer (when evaporation is also higher and the effectiveness of the rain is reduced), the two environments are not too dissimilar (grey and brown soils of heavy texture are found in areas receiving as much as 50″ per annum, but generally less than 40″).

Although grey and brown soils of heavy texture are found in areas of higher rainfall in northern Australia, their morphology greatly reflects the effects of the winter aridity—soluble salts such as sodium chloride accumulate in the profile, and gypsum, which is more easily removed from the soil profiles than lime, accumulates in the lower horizons. Lime is less important in these soils than gypsum, possibly because the alluvium from which both are derived is generally gypseous.

The grey and brown soils of heavy texture are divided into two separate Great Soil Groups on the basis of their colour—the grey soils of heavy texture and the brown soils of heavy texture. There is merit in the inclusion by Stace, Hubble and others (1968) of a third Great Soil Group with these soils, namely the reddish, heavy textured desert soils, as not all soils on heavy textured alluvium in dry areas are either grey or brown. The heavy textured alluvial soils north of Broken Hill, in New South Wales, for example, are certainly reddish-brown, rather than grey-brown.

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[4] This name is used by Stephens (1962). Stace, Hubble and others (1968) include these soils with grey, brown and red clays.

Location: Northeastern corner of the Deniboota Irrigation District, 3 miles from the Edward River, Riverina, New South Wales.							**Vegetation and Land-use:** originally grassland with saltbush, now irrigated farming.			
Drainage: very poor.		**Parent Material:** heavy textured alluvium.					**Slope:** flat.	**Average Annual Rainfall:** 15".		
Horizon	**Depth**	**Lower Boundary**	**Colour**	**Moisture**	**Texture**	**Structure**	**Lime ($CaCO_3$) and Gypsum ($CaSO_4 . 2H_2O$)**	**pH**	**Special Features**	
A	0"-1"	gradational, irregular	light grey	dry	medium clay	apedal massive	nil	7.0	very hard	Chlorides present in all horizons
B1	1"-10"		dark grey		heavy clay	weak coarse blocky	nil	8.0	hard	
Bca	10"-30"		yellowish-grey			apedal massive	trace of fine earth lime, no gypsum	7.5		
Bcs	30"-50"		mid grey				no lime, but coarse gypsum crystals decreasing with depth	7.5		
C	50"+	heavy textured alluvium containing some gypsum								

Figure 10.9: Field profile description of a grey soil of heavy texture from the Riverina, New South Wales (modified from E. J. Johnston, 1953).

In Fig. 10.9 a grey soil of heavy texture from the Riverina in New South Wales is described. Note the lack of distinct horizons—the entire solum is greyish heavy clay. The *A* horizon consists of a thin, hard covering to the soil and makes water penetration extremely difficult. The profile is poorly structured because the clay is sodium-dominated and therefore dispersed; in the dry season the clay sets almost like concrete. Experiments on water penetration into these soils have shown that sometimes it takes several days for water to penetrate a few centimetres. Lime and gypsum are absent from the *A* and *B* horizons, but some lime accumulates in the *Bca* horizon (there is hardly enough lime to warrant the suffix *ca*) and large quantities of gypsum are present in the *Bcs* horizon immediately overlying the parent alluvium, which is too deep to become organised into solum, but still contains gypsum and other salts (its *pH* is also about 7.5). The *pH* of this soil is related to the free soluble salts which are close to neutral in reaction, being derived from strong acids and strong bases (e.g., sodium chloride and gypsum, calcium sulphate), unlike the *pH* of chernozem subsoils which is related to the presence of lime (a salt from a strong base and a weak acid, so the *pH* is higher).

Horizon	Percentage Clay	Percentage Total Soluble Salts	Total Chlorides (as NaCl)	
A	46	0.045	0.021	
B	60	0.412	0.250	
Bca	59-61	0.505-0.845	0.297-0.550	the first figure refers to the top of the horizon and the second to the bottom
Bcs	58-67	1.176-0.511	0.505-0.306	
C	27	0.915	0.441	

Figure 10.10: Percentage clay, total soluble salts, and chlorides (as sodium chloride) in the grey soil of heavy texture described in Figure 11.8 (after E. J. Johnston).

The processes operating in this soil (Fig. 10.9) are: clay accumulation in all horizons (Fig. 10.10 shows the clay content in all horizons, with the exception of the hard crust, to be about 60%, compared with 27% for the underlying alluvium); illuviation of gypsum and some lime; and salt accumulation in all horizons, with a concentration in the *Bcs* horizon (Fig. 10.10 shows that total soluble salts and chlorides, while at maximum concentration in the *Bcs* horizon, are present throughout the profile, being higher in the underlying alluvium than in the *A* or *B* horizons). Humus does not feature at all in this soil in the field. Analyses have shown that organic matter is usually less than 3%

in the thin *A* horizon and greatly diminishes in the underlying horizons, to less than 0.5% in the *Bca* horizon. Some brown soils of heavy texture contain slightly more organic matter.

Subsoil mottling is a common feature of these soils, reflecting the periodic inundation and desiccation. The mineralogy of the clay fraction also reflects these conditions—typically, the hydrous micas, especially illite, are featured. Because of the low weathering intensity and presence of bases, 2:1 rather than 1:1 lattice clays dominate. Appreciable quantities of 1:1 lattice clays such as kaolinite are more likely to be the result of inheritance than of formation *in situ*.

The clayey parent material and the low relief, resulting in periodic inundation, are the two soil-forming factors which dominate the morphology of grey and brown soils of heavy texture. Sandy parent materials in well-drained situations form other Great Soil Groups such as red-brown earths (see Chapter 11) which frequently form catenas with grey and brown soils of heavy texture. Since these soils are found within a wide rainfall range, climate is less limiting to their occurrence than relief or parent material which combine to reduce leaching effectiveness even in the wetter areas (more than 30″ average annual rainfall). However, the sporadic nature of the rainfall is an additional factor contributing to the ineffective leaching. These soils, then, appear to be more intrazonal than zonal in character, and their widespread occurrence in Australia, which has led pedologists to consider them as zonal, may be accounted for by the coincidental deposition of fine textured alluvium over vast low-lying areas which are subject to inundation.

Grey and brown soils of heavy texture, because of their large amount of swelling clay, like the black earths and chernozems, are also very difficult to manage after rain; roads across them (such as those in western Queensland) usually become impassable. They are more difficult to irrigate than black earths and chernozems because of their low permeability and high salt content, but some have been reclaimed in the Riverina where they are suitable for rice growing (they retain water very well) and fodder crops. Elsewhere, they are generally grazed in their native state. Their carrying capacity varies enormously with the weather: after rain, lush tall grasses grow and carrying capacity increases greatly, but during dry spells their carrying capacity is very low. To make use of the excellent grazing which occurs intermittently, attempts are being made, especially in the Channel country, to move stock, after rains, into areas of these soils where they can then be fattened before being sent to the market. Without stock movements, the overall carrying capacity of these soils is quite low as

farmers have to stock for dry spells rather than for good seasons. In western Queensland, where these soils are most widespread, cattle breeding is the main land-use. The plains of the grey and brown soils of heavy texture, like the black soil plains, are generally treeless, but low shrubs, especially salt and blue bush, are quite common.

A very important feature of all the soils considered so far in this chapter is gilgai, which is optimally developed in heavy textured soils high in montmorillonite. Gilgai is a surface undulation, or micro-relief, consisting of ridges (called puffs) and depressions (called shelves) caused by differential swelling and shrinking of the soils. The difference in height between the puffs and shelves may be several feet but is often only about 18 inches. The shelves often contain a hole measuring a few feet in width and a foot or so in depth: this is called a 'melon-hole'. Gilgai forms distinct patterns across paddocks. These patterns are obvious from the air and pedologists often use air photographs to study them. The soil profiles on the puff differ from those on the shelf where the eluvial horizon is usually better-developed. Black earths, chernozems and grey and brown soils of heavy texture are particularly prone to gilgai formation, but it is also quite common in heavy textured soils of sloping areas (such as the chocolate and prairie soils). Water frequently remains in the melon-holes and shelves of grey and brown soils of heavy texture long after the puffs have dried out, further accentuating the effect of the micro-relief.

SOILS ON LIMESTONE

Limestone and other related rocks such as marls (calcareous and siliceous fine-grained rocks) and dolostones (calcium-magnesium carbonate rocks) represent one extreme type of parent material offered for soil formation; the other extremes are quartz sandstones and quartzites, which are as low as limestone is high in bases and clay-forming minerals. Limestones resist many of the formal zonal soil-forming processes such as podzolisation, and in areas of podzolised soils they form intrazonal clayey soils instead. They may form zonal soils where the soil-forming processes are either so intense that the nature of the parent material is rendered subordinate (e.g., laterisation and ferralisation, forming lateritic soils and krasnozems respectively) or so weak that illuviation of clay, humus and iron do not occur (e.g., in sub-humid to semi-arid grasslands). Thus limestones may give rise to zonal soils in the hot, wet tropics and in areas as dry as, or drier than, those of chernozems and black earths. Elsewhere, limestones form intrazonal soils.

Three main types of soils are recognised on limestones (excluding the krasnozem and lateritic soils of the hot, wet tropics); unfortunately, the role of the five factors of soil formation in distinguishing these types is not clearly understood. In wetter areas, terra rossas (from Latin, meaning red earth) are more likely to form; in sub-humid and semi-arid areas, rendzinas; but there is considerable overlap between these two. In desert areas, shallow, lithosolic, grey, brown and red, calcareous desert soils form.

Provided that the relief, type of limestone, vegetation, and duration of weathering are similar, terra rossas form in areas of higher rainfall than rendzinas, but if relief varies and all the other factors, including climate, are similar, terra rossas form on well-drained hillslopes and rendzinas in poorly drained depressions. This occurs in the Ashford area of New South Wales; the profile described in Fig. 10.11 is from this area. In areas with similar environmental factors and durations of weathering, terra rossas tend to form from harder limestones, and rendzinas form from softer limestones such as chalks and marls. If only vegetation varies, terra rossas tend to form under forest vegetation and rendzinas under grassland. In sub-humid areas, terra rossas are frequently the result of weathering in a previous wetter climate and so are relict soils, while rendzinas form from the weathering of fresh limestone. Thus all the factors of soil formation can account for the morphological differences between terra rossas and rendzinas.

So terra rossas are more leached than rendzinas as they are associated with higher rainfall, better drainage and forest vegetation. Most terra rossas have an acid reaction, and, aside from floaters of parent rock, contain no free calcium carbonate in their profile. Since their parent material consists mostly of calcium carbonate, their mineral matter must be derived from the 'impurities' in the limestone; a considerable depth of limestone is required to form only a shallow soil. Rendzinas, on the other hand, generally contain lime throughout their profiles, so though they form by less intense weathering than terra rossas, they are frequently deeper. The grey, brown and red, calcareous desert soils are usually very shallow because of the very weak weathering intensity in deserts.

TERRA ROSSAS

Terra rossas are intrazonal soils on limestone where leaching is moderately intense. They are common in areas containing podzolised soils on less basic rocks. The Great Soil Group most resembling terra rossas is the one—the shallow red soils—

Location: 11 miles northwest of Ashford, on the road to Ashford Caves north of Inverell, New South Wales.

Vegetation and Land-use: natural grass grazed by cattle.

Parent Material and Underlying Rock: hard limestone.

Slope: 4°. Slope has outcropping boulders between the soil.

Drainage: good.

Average Annual Rainfall: 30″.

Horizon	Depth	Lower Boundary	Colour	Moisture	Texture	Structure	Lime	pH	Special Features
A	0″-3″	gradational, irregular	brownish-red	moist	light clay	weak crumb	nil	5.5	hard fragments of limestone up to 10″ across scattered throughout solum
B	3″-20″	very sharp, irregular	mid bright red		medium clay	irregular, subangular, blocky	nil	6.0	
R	20″+	hard, fresh limestone.							

Figure 10.11: Field profile description of a terra rossa in northern New South Wales.

discussed earlier in this chapter. The shallow red soil described in Fig. 10.5 would be identical in morphology to the terra rossas described in Fig. 10.11, but for one important difference: there is no *C* horizon in the terra rossa. Most soils on limestone lack weathering rock zones beneath their sola, which rest sharply on fresh limestone (see Figs. 10.11, 10.12 and 10.13). The shallow red soil described in Fig. 10.5, however, contains quite a deep zone of weathering basalt. Terra rossas, like shallow red soils, differ from krasnozems in their shallow depth and lower sesquioxide accumulation.

The terra rossa described in Fig. 10.11 is typical of the soils formed on the hard Palaeozoic limestones found intermittently along the western slopes from Victoria through to Queensland. These limestones are well-known for their caves, some of which are open to tourists (e.g., Buchan in Victoria, Wellington and Jenolan in New South Wales, and Olsen's near Rockhampton in Queensland). The landforms and soils on all these outcrops are similar, though some do not contain caves (e.g., Mudgee limestone). These limestones represent the edge of the large Silurian sea which covered much of central-eastern Australia. The only situation on these hard limestones which does not give rise to terra rossas is the poorly drained creek areas, especially if the parent material is alluvial rather than rock limestone. The soil described in Fig. 10.11 grades into a greyish, heavy textured soil along creeks. This soil resembles poorly developed rendzinas; the main difference between it and true rendzinas is the drab colouring and lack of humus in the *A* horizon. Softer more recent limestones in the same area give rise to rendzinas such as the one described in Fig. 10.12.

The terra rossa shown in Fig. 10.11 is distinctly red in colour—as red as krasnozems. The upper few inches are slightly darkened by organic matter, which is less than 2% in the *A* horizon and less than 1% in the *B* horizon. Some clay illuviation is apparent in the slight texture change down the profile (the *A* horizon contains 47% clay and the *B* horizon 59%), but clay accumulates in both the *A* and *B* horizons. The soil is fairly well-structured because calcium is the dominating cation on the colloids and free iron oxides cement the flocculated colloids. The solum does not react to hydrochloric acid and the *pH* is weakly acid. Even the soil close to the limestone floaters failed to react to hydrochloric acid, though its *pH* was 6.5. Removal of free calcium carbonate from this soil is therefore very thorough though the soil colloids contain hydrogen together with appreciable exchangeable calcium. In this respect terra rossas differ from shallow red soils which are hydrogen-dominated.

Location: 22 miles west of Toowoomba, Darling Downs, Queensland.
Vegetation and Land-use: formerly grassland, now grain and fodder crops.
Parent Material and Underlying Rock: soft, marly limestone.
Slope: less than 1°.
Drainage: moderately good.
Average annual rainfall: 25.6".

Horizon	Depth	Lower Boundary	Colour	Moisture	Texture	Structure	Lime	pH
A	0"-9"	gradational, irregular	dark brownish-grey	moist	light clay	moderate granular	some fragments, increasing in size and frequency with depth	8.5
B	9"-22"	very sharp, irregular	brownish-grey, some yellow			nutty	many fragments up to 6" across, increasing with depth	8.5
R	22"+	fresh, soft, marly limestone.						

Figure 10.12: Field profile description of a rendzina from the Darling Downs, Queensland (modified from Thompson and Beckman, 1959).

The soil-forming processes in terra rossas are similar to those of shallow red soils: rapid decomposition of humus, resulting in lack of dark colours; accumulation of iron oxides throughout the solum, producing red colours; accumulation of silicate clays in the solum, with some illuviation to the *B* horizon; removal of free calcium carbonate, giving an acid reaction; and the complete removal of alkali cations. The floaters of limestone within the solum are analogous to the floaters of basalt within the solum of shallow red soils.

Because Australian terra rossas are found in areas of fairly rugged relief, they are left mainly for rough grazing, in spite of their inherently high natural fertility. Where they are cultivated, for example, on the softer limestones in the wetter areas near Adelaide, they are very productive.

Rendzinas

The main area of rendzinas in Australia is in the south-east of South Australia not far from Mt. Gambier, where they are found on soft Tertiary limestones, on the drier inland side of terra rossas. They also occur sporadically on the Darling Downs in Queensland, and in other areas of soft limestone in eastern Australia. Their morphology contrasts greatly with that of terra rossas. They bear some similarities to black earths, but the similarities are far less striking than those between terra rossas and shallow red soils.

Rendzinas are incompletely leached soils in which lime accumulates; however, considerable leaching of lime must have occurred to account for the large amount of silicate clay present in rendzinas, compared with their parent materials which are mostly calcium carbonate. The rendzina from the Darling Downs described in Fig. 10.12 is a light clay throughout. There is little profile differentiation, apart from the development of a dark brown colour in the *A* horizon from the accumulation of humus (the *A* horizon contains more than 8% organic matter), and an increase in the amount of lime in the solum as the parent material is approached. The soil is well-structured, from the combined effects of calcium domination and humus. Like the terra rossa described in Fig. 10.11, the rendzina described in Fig. 10.12 also has a sharp boundary with the underlying rock and lacks a *C* horizon. The *pH* throughout the profile is close to that of free calcium carbonate, namely, 8.5, to the nearest 0.5 *pH* units.

The soil-forming processes in rendzinas include: removal of alkali cations; accumulation of lime throughout, with some illuviation to the *B* horizon; accumulation of humus high in humic acids in the *A* horizon, decreasing with depth; and accumulation of silicate clay throughout the solum (no illuviation). Considerable leaching of lime must occur to allow these other processes to function (cf. the grey, brown, and red calcareous desert soils in which these processes do not function because there is insufficient lime removal). In these respects, rendzinas are not unlike chernozems, but their *A* horizons are usually not as black as chernozems, though rendzinas also are associated with native grassland vegetation. Frequently chernozems and rendzinas are found side by side in the sub-humid to semi-arid grasslands. On the Darling Downs, the profile described in Fig. 10.12 is associated with black earths.

Because of their high calcium carbonate content, rendzinas are excellent soils for grapes and are used for vineyards in parts of the famous Barossa Valley in South Australia. They are also suitable for fodder grasses which require a lot of calcium and do not mind an excess. Like black earths, they are very productive soils.

In very poorly drained situations, the lower horizons of rendzinas may be intermittently invaded by the water table, which may cause some mottling analogous to that in the grey and brown soils of heavy texture. This invasion by ground waters may introduce salts into the rendzina profile and make them less productive. Such soils are called *ground water rendzinas*.

Grey, Brown and Red Calcareous Desert Soils

In the strict sense, the grey, brown and red calcareous desert soils are not soils at all, for they do not contain genetically related horizons: they are really lithosols on desert limestones. However, some have incipient soil characteristics such as less lime in the upper few inches. Generally, they represent disintegrated limestone rubble in which little leaching has occurred. These soils are common in the Australian Arid Region in all areas of calcareous rock, including calcareous sandstones and shales, as well as limestones such as the large outcrop on the Nullarbor Plain in southern Australia.

This lack of pedological organisation is exemplified by the profile from north of Broken Hill in New South Wales described in Fig. 10.13. The profile consists of a shallow, loamy, calcareous solum, resting abruptly on fresh limestone. There is little evidence of any illuviation of lime as it accumulates even on the soil surface. No peds are present and the *pH* is 8.5, strengthening the suggestion that little change has taken place beyond chemical and physical decomposition. The solum is reddish-brown and there is no evidence of humus accumulation. It appears, then, that horizon-differentiating processes are absent or minimal.

Horizon	Depth	Lower Boundary	Colour	Moisture	Texture	Structure	Lime	pH	Special Features
Location: Fowler's Gap Research Station, 70 miles north of Broken Hill, New South Wales.						**Vegetation and Land-use:** desert shrubs and sparse grass, grazed by sheep.			
Parent Material and Underlying Rock: hard limestone.						**Slope:** 6°, near a rocky outcrop.			
Drainage: good.						**Average Annual Rainfall:** 9".			
A	0"-9"	sharp, irregular	light reddish-brown	dry	loam	apedal, single-grained	much fine earth and rock fragments scattered throughout	8.5	shallow, stony soil formed between limestone outcrops; lime lies on soil surface
R	9"+	fresh limestone							

Figure 10.13: Field profile description of a reddish-brown, calcareous desert soil from near Broken Hill, New South Wales.

The very low rainfall accounts for the lack of removal of lime from these soils. This accumulation impedes the other processes of soil formation. The colour of these soils is closely related to the impurities within the limestone, which are also not removed from the profile. Some soils are red, some grey, and others brown, depending on the nature and presence of iron oxides.

These soils have found little use because of the very low rainfalls. They are used mainly for rough grazing, but providing water for the stock is a major problem.

The soils discussed in this chapter have a common process: the accumulation of silicate clays and lack of, or minimal, texture-differentiation. They include pedalfers and pedocals, all of which have certain limitations with respect to parent material. The next chapter will discuss soils in which lime or salts accumulate, that is, the soils of arid and semi-arid regions, though some of the soils of these regions have already been considered in this chapter.

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Chapter Eleven

Soils Dominated by Accumulations of Bases

CONTENTS

THE TERM 'BASE' HAS BEEN USED IN THIS BOOK FOR THE CATIONS of the alkali metals (especially sodium and potassium) and alkaline earths (especially calcium and to a lesser extent magnesium). Because basic cations are so mobile and readily tend to form soluble compounds such as sodium chloride in the soil profile, they are swiftly removed from soils of humid areas. Non-basic cations such as iron and aluminium are less mobile and tend to form very insoluble compounds such as hydrated oxides, which strongly resist leaching. In dry areas or areas of poor drainage, such as topographic depressions, bases are able to accumulate within the soil profile because of ineffective leaching. Thus base-accumulating soils occur zonally in arid and semi-arid regions and intrazonally in the depressed areas of more humid regions.

The soils dealt with in this chapter have in common the accumulation of bases somewhere in their profiles: the bases may be found in all horizons, but more commonly they are concentrated in certain horizons by illuviation, and less frequently by capillary rise. Some horizons may even be base-depleted, especially in semi-arid areas, the bases accumulating only in the lower horizons of the profile. The compounds formed from bases in the soil have varying solubilities, so some (e.g., calcium carbonate) tend to accumulate higher in the profile than others (e.g., calcium sulphate). The alkali cations form more soluble compounds than the alkaline earths and so are restricted more to drier, poorly leached areas. Compounds formed from alkaline earths are found in both arid and semi-arid areas.

Bases accumulate in soils in two principal forms. First, they may be found as free salts, that is, in combination with soil anions; as free salts (e.g., sodium chloride, calcium carbonate) they contribute to the mineral matter of the soil. Second, they may be found as cations adsorbed to the soil colloids, especially in the case of silicate clays. In humid areas, hydrogen is the dominating cation on soil colloids, but in arid and semi-arid areas, much of the hydrogen is replaced by basic cations because they are more highly concentrated in the soil solution. Thus the colloids of soils of dry areas are base, rather than hydrogen, dominated. Bases do not exist freely as salts in the soil until the colloids are base-dominated. Once this condition is fulfilled and if there are still some bases remaining unadsorbed by the colloids, free salts may form; that is, the presence of free salts of a certain base indicates a concentration in the soil of that base in excess of what is required to exchange the cations on the soil colloids. Bases as free salts are easily identified in the field by their white colour, salty taste and reaction to silver nitrate (see Chapter 2); but bases on the colloid are more difficult to detect unless their presence is reflected in the soil's *pH* (sodium and potassium on the colloid raise the *pH* above 8.5; soils with calcium-dominated colloids usually have *pH* values close to 8.5), so many pedologists analyse the soil's exchangeable cations to measure these bases.

Whether the bases exist as free salts or only as cations on the soil colloids is vital to soil morphology, especially to soil structure. Cations existing as free salts in the soil may have the opposite effect on flocculation and dispersion to that which occurs with cations dominating the soil colloids; for example, soils with sodium-dominated colloids but no free sodium salts are dispersed and lack peds, whereas soils with free sodium salts in addition to sodium-dominated colloids are flocculated and so have better ped formation. Similar effects obtain with calcium: soils with calcium-dominated colloids are flocculated and well-structured, but too much free calcium present as salts, such as calcium carbonate, causes dispersion.

The morphology of the soils studied in this chapter, then, is related in no small way to the nature, concentration and distribution of bases within the profile. The role of bases in the soil also influences the other soil-forming processes which might occur; for example, illuviation of clay is related to the state of flocculation and dispersion of the soil, and this, in turn, is controlled by the base status of the soil. The relative importance of the role of bases decreases, and the relative importance of the other processes of soil formation increases, with decreasing aridity. Since this chapter includes the examination of soils of semi-arid areas, the role of bases is not as important for some of the soils discussed, with the soil-forming processes found in the soils of humid areas being just as important. For example, in the red-brown earths, bases dominate only in the subsoil, and the surface horizons have a closer affinity with the soils of humid areas than with those of arid areas.

The usage of the terms relating to base accumulation in this book requires explanation, since the terms are used differently in the various disciplines. The term *salt*, in chemistry, refers to any combination of alkali cation with acid anion (base with acid), irrespective of whether sodium is involved. In pedology, however, the term *salt* is more frequently associated with sodium combinations, e.g., sodium chloride, sodium carbonate, sodium sulphate. The general term *salty soil* implies the presence of sodium in some form. There are two main types of salty soils, saline and alkaline. *Saline soils* contain free sodium salts as well as sodium-dominated colloids; their *pH* is usually not high, for the free salts are commonly neutral salts (such as sodium chloride) which have a *pH* of 7. *Sodic or alkaline soils*, on the other hand, contain no free sodium salts but their colloids are sodium-dominated: they therefore have a high *pH* (quite often 9.5 or 10.0). Both types of salty soil (saline, and alkaline or sodic) contain sufficient sodium to interfere with the growth of most crops, saline soils being chemically poorer (they frequently contain large quantities of chloride which is particularly harmful to plant growth) but physically better than alkaline soils (the free sodium salts cause flocculation rather than dispersion). Both saline and alkaline (sodic) soils present serious problems for cultivation. *Calcareous soils* (soils dominated by calcium, either on the colloids or as free carbonates), do not present such serious cultivation problems, because calcium, besides being an essential plant nutrient, combines with harmless carbonate (CO_3) rather than harmful chloride (Cl).

Any soil can become saline, alkaline or calcareous by the accession of the appropriate bases under conditions which prevent their removal by leaching. Frequently, however, bases accumulate in the soils of dry areas by incomplete leaching of the parent material, but the parent material is only one of many sources of the bases found in soils. The origin of the bases accumulating in soils may be listed as follows:

1. The bases may be inherited directly from the parent material under ineffective leaching. In rock-derived soils, bases may derive from the feldspars which contain sodium, calcium and potassium in their structures, and from other aluminosilicate minerals whose tetrahedra (see Chapter 4) are linked by basic cations. In soils

derived from some sedimentary rocks and pre-weathered materials, bases may be present (in addition to the aluminosilicate minerals) as free salts (e.g., in calcareous shales) or as cations attached to the colloidal minerals.

2. In coastal areas, bases may be added to the upper soil horizons directly from sea spray or swash action associated with high waves at full tides. Invasion of coastal soil profiles by saline groundwater may add bases to the soil from the lower horizons. If rainfall is high, these bases are soon leached, but in dry coastal situations, they may accumulate to form very saline soils. Since sodium is one of the most soluble of all cations, it is not surprising that it dominates sea-water which builds up its salinity over the years by receiving river water which has at some time leached salts from soil profiles. Sea-water contributes mainly sodium to the soils it invades, but other cations are also important.

Salt-bearing groundwaters are by no means restricted to coastal locations: many inland groundwaters are also quite saline and may invade soil profiles naturally, or over-irrigation may raise the water table of an area to invade the soil profile. All irrigation water and groundwater contains some minerals, but the quality and quantity are variable. The actual damage done to soils by these waters also depends on the concentration of salts within the soil itself, so that the same water may seriously damage one soil and not another. If the water is known to be damaging to the soil it invades, gypsum ($CaSO_4 \cdot 2H_2O$) may be added to reduce the amount of sodium absorbed by the soil, or the water may be demineralised before it is applied.

3. Many bases are added to the soil annually as *cyclic salt*, the term given to the bases added to the soil by rainwater. By its taste and domestic use as 'tank water' (rainwater collected from roofs), rainwater would appear to be low in bases. However, small quantities of bases are present, especially sodium, calcium and potassium, and, with ineffective leaching, even such low concentrations form appreciable accumulations within the soil. In areas with an average annual rainfall of more than 15″, the accumulation of bases from cyclic salt is negligible, for leaching removes them. But in areas where the rainfall is less than this, the accumulation may be considerable.

Recently, in Australia, cyclic salt has received considerable attention in the study of the soils of arid and semi-arid areas. The origin of the bases added to the soil as cyclic salt is not fully understood. Some undoubtedly comes from the ocean itself (low cloud passing across the coast could collect sea spray) but the ratios between the various bases in rainwater do not agree with the ratios in seawater. Also, the ratios between the various bases

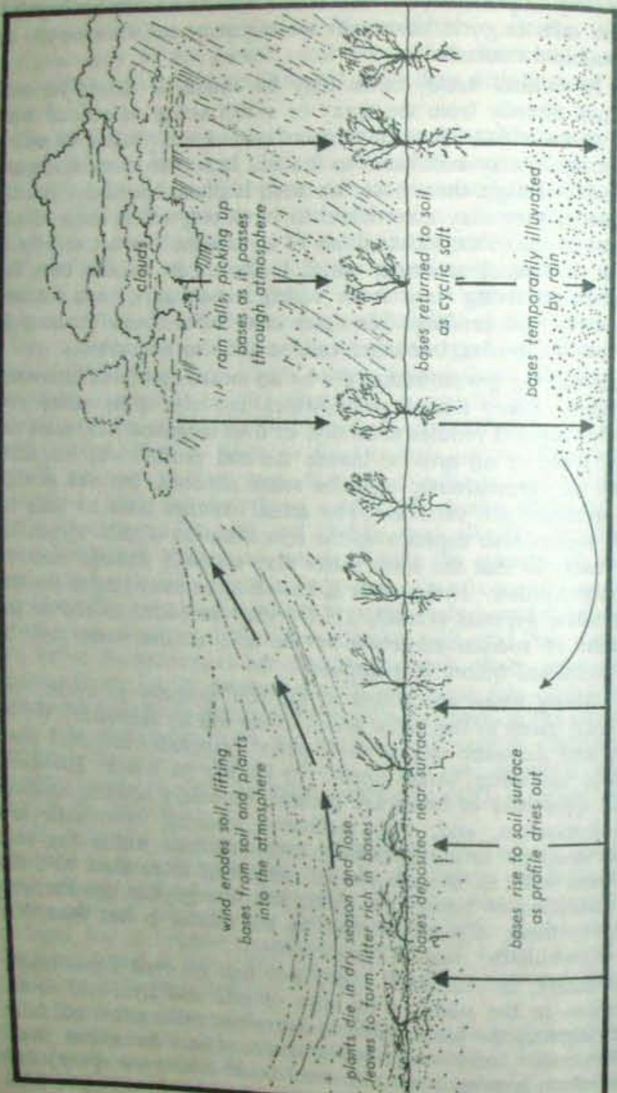

Figure 11.1: Cycling of bases from plants and soil into atmosphere to form cyclic salt.

in rainwater vary with the seasons, and from site to site (if the ocean was their sole source this variation should not occur). Hutton (1968) believes that most of the bases in rainwater are acquired as the rain falls through the aeolian dust and plant remains in the atmosphere. When soils dry out, bases frequently concentrate at or near the surface by deposition as the soil water evaporates. Wind erosion removes these bases with the topsoil and they become part of the aeolian dust frequently found in dry areas. Thus there is a cycle whereby bases rise to the soil surface in dry periods, are taken into the atmosphere by wind erosion to become cyclic salt, and in this form are later returned to the soil with the rain, perhaps in some other area. Rainwater illuviates some of the bases to the subsoil, but when the profile dries out, the bases rise and the cycle is repeated. The cycle is shown in Fig. 11.1.

Thus there are several possible sources for the bases found in the soils of dry areas, and, because of their mobility there is a very delicate balance between their removal and accumulation. The balance depends on rainfall and site. Since bases are so soluble, they move through the soil profile with the soil water: they are lower in the soil after rain and higher in the soil in dry periods.

Plants also assist in the cycling of bases through the soil; for example, salt-tolerant plants such as salt and blue bush return large amounts of bases to the soil surface. If such plants are eaten by animals or die from drought, the bases they have returned to the soil surface, together with their remains, may be wind eroded and contribute to the base content of cyclic salt.

Thus bases play a very dynamic role in the soils of dry areas, and the movement of bases through soils is perhaps the most rapid of the soil-forming processes: a single shower of rain could temporarily illuviate some of the desert soil's bases and cause an appreciable change in the *pH*, the total soluble salts, and the exchangeable cations down the profile. In humid areas, a single shower of rain would be unlikely to cause any measurable changes in soil properties, such as an increase in clay content in the subsoil (from the clay illuviated by the rainwater) or *pH*. Thus, bases, unlike other soil constituents, are very mobile, and one of our chief concerns in this chapter is to study their movements through the soil and their replacement (natural or artificial) by other bases and hydrogen.

When studying the field profile descriptions in this chapter, the student needs to realise that the properties relating to bases (*pH*, depth, and amount of lime and gypsum) may vary from season to season, especially in the soils of the deserts, and that cultivation, especially if salt removal is necessary, drastically alters these properties.

There are many Great Soil Groups that could be included here if we were to discuss all the soils of arid and semi-arid areas. However, to simplify our study and to demonstrate the interaction of the soil-forming processes, only the most important soils are considered. Polygenetic arid and semi-arid soils, such as the silcreted, deeply weathered soils, calcareous red earths, red and brown hardpan soils, brown soils of light texture, and solodic soils are given more attention in Chapter 12 (relict soils) while some are only briefly considered here. Some of the soils of arid and semi-arid areas have already been discussed in Chapter 10 (soils dominated by silicate clays), namely the grey and brown soils of heavy texture and the grey, brown, and red calcareous desert soils, since these soils form part of successions of soils on heavy-textured parent materials, though they also accumulate bases in their profiles and could therefore be included here. Thus, information on the soils of arid and semi-arid regions is contained in Chapters 10 and 12 as well as in this chapter.

The Great Soil Groups considered here are: the red-brown earths and brown earths (zonal soils of semi-arid areas—soils in which lime rather than salt is more important); solonised brown soils (also called mallee soils, these too are zonal soils on transported parent materials in wetter arid and drier semi-arid areas); the succession of soils representing various stages in salt accumulation and removal (solonchak, solonetz, solodised solonetz and solodic soils); and some of the poorly differentiated desert soils (desert loams, desert sands and sierozems).

RED-BROWN EARTHS

Red-brown earths[1] are the most common soil of Australian semi-arid to sub-humid areas. In southern Australia they dominate in areas between the 14″ to 25″ average annual isohyets, and in northern Australia they are sporadically distributed in areas having an average annual rainfall of 20″ to 35″. Their distribution in southern Australia coincides closely with the wheat belt, so they are one of the most important soils for cultivation and for this reason have received a great deal of attention. They occur on a wide variety of substrates: acid and intermediate igneous rocks, most sedimentary and metamorphic rocks, sandy alluvium, wind-transported material—in fact, on most parent materials, but not the very heavy textured ones which form black earths or grey and brown soils of heavy texture instead. Red-brown earths require free drainage (if drainage is poor, wiesenbodens or grey and brown soils of heavy texture form) and are associated with natural sclerophyll woodland vegetation. They are therefore zonal soils, responding to climate and free drainage.

A typical red-brown earth profile is described in Fig. 11.2. Note the shallow *A* horizon underlain by a very poorly developed *E* horizon, distinguished solely on the basis of its lack of brown colour which, in the *A* horizon, results from humus. The *E* horizon is so poorly developed that the use of the letter *E* is questionable, but in other red-brown earth profiles, bleaching does occur; such bleaching is thought to be related to incipient podzolisation rather than to the processes concerned with red-brown earth formation. The *Bt* horizon is heavier textured than the eluvial horizons, suggesting clay illuviation, though the importance of this process in causing the texture contrast has recently been disputed by Oertel (1961 and 1968) who claims that the higher clay content of the *B* horizon is more the result of *in situ* weathering (if this is so, the *Bt* horizon should be labelled *Bv*) and that clay illuviation is relatively unimportant. It is probably safe to assume that some clay illuviation has occurred, though it may not account for all the texture contrast between the eluvial and illuvial horizons. Maximum clay content is in the *Bt* horizon and decreases with depth.

The eluvial and *Bt* horizons described in Fig. 11.2 have been depleted of free lime. The eluvial horizons are sufficiently leached to have *pH* values below 7, but there is still considerable calcium on the clay minerals. There is usually a sharp rise in the amount of exchangeable calcium in the *B* horizons, especially in the *Bca* horizon in which free lime accumulates. Gypsum, being more soluble than lime, is carried further down the profile and accumulates beneath the *Bca* horizon (in the *Bcs* horizon). The succession of a lime accumulation overlying a gypsum accumulation in soils is a common one (see also Fig. 11.3). The *pH* of both the *Bca* and *Bcs* horizons is 9, suggesting an increase in alkali cations on the colloids.

The parent material of the profile described in Fig. 11.2 is well-drained alluvium of the prior stream systems of the Murrum-

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[1] The name red-brown earth is used for Australian soils which are similar morphologically to American reddish-chestnut soils. Australian brown earths resemble American chestnut soils. The name red-brown earth is not widely used outside Australia, unlike most of the Great Soil Group names used so far (e.g., podzols, chernozems, lateritic soils, krasnozems which are used in many countries). The terms used for some of the other soils in this chapter favour Autsralian usage, e.g., solonised brown soils, desert loams, but the names of the saline and alkaline soils (solonchak, solonetz, solodised solonetz) are common terms in overseas soils literature.

Location, Deniboota Irrigation District, near the Wakool River, Riverina, New South Wales.

Vegetation and Land-use: stunted savanna woodland, irrigated for grazing on improved pasture.

Parent Material: prior stream alluvium.

Drainage: well-drained, sloping site.

Underlying Rock: not exposed.

Average Annual Rainfall: 16″.

Horizon	Depth	Lower Boundary	Colour	Texture	Structure	Lime	pH	Special Features
A	0″-5″	sharp, regular	light brown	sandy loam	weak crumb	nil	6.0	No gypsum present
E	5″-10″		reddish-brown				6.5	
Bt	10″-18″		mid bright red	medium clay	moderate prismatic		7.5	
Bca	18″-25″	gradational, very irregular	reddish-brown mottled with yellow and grey		weak irregular angular blocky	fine earth throughout, some nodules	9.0	
Bcs	25″-32″	gradational, irregular	yellowish-brown mottled with grey	light clay	apedal massive	trace of fine earth	9.0	gypsum accumulation throughout, decreasing with depth
C	32″+		yellowish-grey mottled with reddish-brown, brown and black	sandy clay loam	alluvial bedding apparent	nil	8.5	texture becomes sandier with depth: a little gypsum in top few inches, none in the rest

Figure 11.2: Field profile description of a red-brown earth from the Deniboota Irrigation District, New South Wales (modified from E. J. Johnston, 1953).

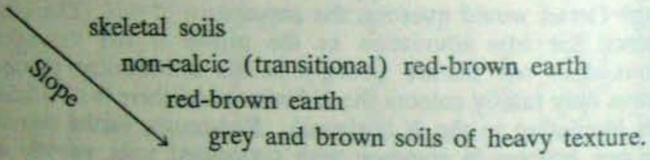
bidgee River (see Chapter 12). This alluvium is variable in texture, but, being close to the prior stream itself, has a high sand component. The red-brown earth developed from this alluvium contrasts markedly in organisation with the alluvium: the soil is pedal (*Bt* horizons of red-brown earths, like the one described in Fig. 11.2, are frequently prismatic in structure, while the underlying *B* horizons are blocky, their pedality diminishing as the *C* horizon is approached) with marked eluvial and illuvial horizons, while the parent alluvium contains bedding planes, and alters in texture relative to these planes. Lime is spasmodic in the alluvium (there is none in the alluvium beneath the soil described in Fig. 11.2) and is not concentrated neatly into one of the *B* horizons as it is in the soil.

From Fig. 11.2, the processes giving rise to red-brown earth formation can be ascertained. Removal of alkali cations is fairly thorough until the lower *B* horizons, where the *pH* rises sharply. Alkaline earths are removed from the eluvial horizons, but free lime is deposited in the middle *B* horizon, and free gypsum in the lower *B* horizon: illuviation of calcium has therefore occurred. Silicate clay illuviation is apparent in the sharp texture contrast, though Oertel would question the importance of this. There is no evidence for iron illuviation as the profile is red throughout. Humus does not feature strongly in the soil-forming processes (humus only faintly colours the *A* horizon and there is no evidence of its illuviation to the *B* horizons). Red-brown earths therefore have one process in common with podzolised soils, namely clay illuviation. Their other processes contrast markedly with those of podzolised soils: they are not base-depleted, iron and humus are not illuviated, and there is no surface accumulation of organic matter. Thus a well-differentiated soil profile can be formed by processes other than podzolisation and laterisation; for example, the horizons of red-brown earths are usually quite distinct, especially with respect to structure, texture, *pH*, and lime and gypsum accumulation, if not colour. There are usually quite sharp boundaries between the horizons (see Fig. 11.2). Red-brown earths are perhaps the most strongly differentiated soils of the drier regions of the world.

The rather low weathering intensity in red-brown earths is reflected in their clay mineral content: they tend to be dominated by illite (a 2:1 lattice clay of the hydrous mica group: it is rapidly destroyed under weathering as intense as that conducive to podzolisation). Lesser amounts of kaolinite (a 1:1 lattice mineral) and iron oxides make up the colloidal fraction of red-brown earths.

As wetter areas are encountered, red-brown earths grade into soils with stronger base depletion, and with more characteristics of podzolisation, such as better development of an *E* horizon. These transitional red-brown earths are common along the western slopes of New South Wales. Near Camden, about 45 miles west-south-west of Sydney, profiles resembling red-brown earths have been described. This area has an average annual rainfall of 29″, so the soils are slightly more leached than typical red-brown earths, and the meagre lime accumulation is quite low in the profile (gypsum does not accumulate at all). Red-brown earths grade into desert loams as drier areas are encountered, rainfall being insufficient to bring about the texture, structure and *pH* contrasts found in typical red-brown earths.

One of the most common catenas associated with red-brown earths is skeletal soils on the upper slope: non-calcic (or transitional) red-brown earths in which lime is either absent or very deep in the profile, below the skeletal soils; red-brown earths along most of the slope; and grey or brown soils of heavy texture on the poorly drained flats on heavier-textured parent material as shown:



Unless the upper slope is fairly steep, the two upper soils may be absent from the catena, so that red-brown earths occupy the entire slope, and there are grey and brown soils of heavy texture on the flats. This type of catena is found on both present and prior stream alluvium in the Riverina. The well-drained sandy levees and backslopes form red-brown earths and the clayey flood-plains form grey and brown soils of heavy texture.

Red-brown earths are not naturally high in fertility, for both physical and chemical reasons. The structure of their upper horizons tends to break down after cultivation, as organic matter is lacking, and there is not sufficient calcium on the colloids to maintain flocculation. The soils tend therefore to set hard after rains, rather than return to their natural weak crumb structure. Their low humus content means that the natural reserves of nitrogen and phosphorus are low, so crops have to be rotated with legumes (to correct nitrogen deficiency), and the addition of vast quantities of superphosphate is needed (especially for wheat). The *pH* of their upper horizons is usually satisfactory, and superphosphate does not lower soil *pH*; thus, they are slightly better off as regards plant nutrients than are podzolised soils. Their light textured eluvial horizons make them very susceptible to sheet, gully and wind erosion, so it is not surprising that they have received a great deal of attention from soil conservationists.

NON-CALCIC BROWN EARTHS

These soils resemble the non-calcic (transitional) red-brown earths except that red colours are either lacking or subdued by the higher humus content. They generally lack an *E* horizon, and neither free lime nor gypsum accumulate within their profile. They are less common in Australian than red-brown earths, but are quite common in North America where they are called chestnut soils. An analogy may be made here with the low frequency of brown podzolics in Australia compared with North America. The reasons for the analogous situations with these two Great Soil Groups could be similar—Australian vegetation is not conducive to the type of litter found in brown podzolics and chestnut soils (brown earths).

Brown earths usually contain a single loamy *A* horizon less than 8 inches thick over a clayey, nutty-structured *B* horizon up to 12 inches deep. Organic matter is maximum in the *A* horizon and imparts a brownish colour, the *B* horizon being reddish-brown. Lime is removed from the solum but may be found in the *C* horizon.

The lack of redness, and the high humus content of brown earths suggest that they form in cooler climates than red-brown earths; in some cases this appears to be so, for they occur in the drier eastern parts of Tasmania, in areas receiving less than 24″ average annual rainfall. However, they are recognised elsewhere in Australia, even in Queensland, on intermediate to basic rocks in well-drained middle to upper slopes where they appear to be intrazonal rather than zonal, for they are found in relatively small occurrences. In properties they appear to be intermediate between red-brown earths and brown podzolics, being more leached than the former and less leached than the latter.

SOLONISED BROWN SOILS

The solonised brown soils (also called mallee soils or mallisols) are found on the drier side of red-brown earths in Australia, generally in areas with less than 15″ average annual rainfall, but sometimes as high as 17″ and usually more than 8″. They are generally associated with transported parent materials such as dune sands, alluvium, loess and lacustrine deposits, but are also found on calcareous rocks, although in such cases aeolian or

Location: Fowler's Gap Research Station, 70 miles north of Broken Hill, New South Wales.

Vegetation and Land-use: scattered salt bush and lower shrubs grazed in unimproved state by sheep.

Parent Material: partly transported deposit on slate.

Slope: even 1°.

Drainage: well-drained.

Underlying Rock: slate.

Average annual rainfall: 9".

Horizon	Depth	Lower Boundary	Colour	Texture	Structure	Lime	pH	Special Features
A1 (crust)	0"-½"	sharp, regular	light yellowish-brown	loam	apedal massive	strong effervescence to HCl, but lime not visible	8.5	scattered quartz stones on surface
A2	½"-14"	sharp, irregular	light brown		apedal single grain		9.0	
K	14"-36"	gradational, irregular	light brown and white	silty loam	apedal massive	hardpan of mostly lime (55%)	10.0	
Bca	36"-48"		light brown		apedal single grain	fine earth throughout	9.5	no visible gypsum
Bca, cs	48"-56"						8.5	gypsum throughout, decreasing with depth
C	56"-78"+		light brown mottled with grey		slate structure preserved	weak effervescence to HCl, no visible lime	8.5	no visible gypsum: weathering slate fragments dominate

Figure 11.3: Field profile description of a solonised brown soil (mallee soil), north of Broken Hill, New South Wales.

alluvial accession cannot be precluded (the profile on slate described in Fig. 11.3 undoubtedly contains some transported material from the quartz stones scattered on the surface of the soil).

Solonised brown soils vary greatly in properties, depending especially on the texture of the parent material, but they all contain a very high concentration of carbonates (calcium carbonate with lesser amounts of magnesium carbonate) usually from one to two feet below the surface. In some solonised brown soils, clay illuviation has occurred and there is some texture change above the lime horizon. High *pH* values in the lower horizons are a common feature of these soils (hence the term solonised in their name). The *pH* of the upper horizons is usually alkaline and the subsoil strongly alkaline (the *pH* values of the soil described in Fig. 11.3 can be considered typical) because sodium features strongly on the colloids. Organic matter is very low and plays little part in soil formation.

Some solonised brown soils, such as the one described in Fig. 11.3, contain appreciable lime throughout their profile, but others contain little or no lime in the upper horizons, all of it being illuviated to the *K* and lower horizons. The nature of the lime accumulations is also variable, but is usually greatest at the top of the *K* horizon, decreasing in the lower horizons. Some *K* horizons consist of cemented nodules, whereas others consist of hardpans of massive lime, as in the soil described in Fig. 11.3. Gypsum may accumulate in the lowest solum horizon before the *C* horizon (see Fig. 11.3). In Fig. 11.3, solonisation[2] is maximum in the *K* and *Bca* horizons and decreases on either side of these horizons. All the horizons contain more than 5% lime, except for the *C* horizon. A thin crust develops on top of this soil and there is a sharp junction between the *A2* and *K* horizons; these features occur in most solonised brown soils.

The amount of carbonate accumulating in solonised brown soils is extremely high (55% in the soil described in Fig. 11.3) compared with that in other zonal soils nearby. Several explanations have been offered for this high accumulation. Although the parent materials of these soils usually contain some carbonate, many pedologists favour an aeolian origin for some of the carbonate in these soils; the grey, brown and red calcareous desert soils of the Nullarbor Plain (which lies due west of the main area of solonised brown soils) have been suggested as the main source of carbonate. Since the solonised brown soils themselves are frequently apedal and light textured near the surface they are

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[2] The term 'solonisation' is explained later in this chapter.

susceptible to wind erosion, probably after the lime has been illuviated from them into the *K* horizon. Thus recurrent accessions and removals of parent material concentrate the lime lower in the profile. A second explanation of the lime accumulation is that some of it was deposited fluvially in areas of internal drainage, and soil formation has since re-distributed it. A third explanation which applies to solonised brown soils on the lee-side of playas is that the lime and salt accession is carried by the prevailing winds from these playas. This explanation seems to account for some of the lime and sodium in solonised brown soils in Western Australia.

The main area of solonised brown soils in Australia is in south-western New South Wales, northwestern Victoria and eastern South Australia, covering the lower Murray - Darling Basins. Other areas border the Nullarbor Plain and occur spasmodically in southern South Australia. Much smaller areas are found in Western Australia. Thus, these soils are confined to southern Australia between the 8″ and 17″ average annual isohyets. They are associated with mallee vegetation, especially in New South Wales. The mallee tree is a eucalypt which grows much taller than other trees in neighbouring low rainfall areas. It contains a large bole beneath the soil to store water for dry periods and is extremely tolerant of the high lime and sodium content of the solonised brown soils. The mallee tree is often the only vegetation found in areas of this soil; grasses, aside from tussock, are usually scarce. In their natural state, areas of solonised brown soils offer little for either cultivation or grazing. The mallee trees have first to be cleared, and removal of the bole is extremely expensive (on some farms, the bole has been left and the 'stump-jump' plough is used). Some solonised brown soils are too alkaline for cultivation and others lack trace elements, especially manganese and zinc. The nitrogen and phosphorus reserves of these soils are also low, so rotation with legumes and application of superphosphate are necessary. Water shortage is perhaps the most serious factor limiting the cultivation of these soils. In areas too dry for wheat (less than 12″ average annual rainfall) these soils are used for sheep grazing. In the Murray Valley they are irrigated for fruits, grapes and lucerne, but salting in the lower sites poses problems.

SALINE AND ALKALINE SOILS

Solonised brown soils are affected by excess sodium ions in their subsoils and therefore may be considered as alkaline soils; however, their lime pan (*K* horizon) is the most spectacular feature of their morphology, so they are treated as a separate group rather than included with the soils mentioned here. Similarly, the grey and brown soils of heavy texture could also be included as saline or alkaline soils.

The succession of soils considered in this section (solonchak, solonetz, solodised solonetz and solodic) illustrate the delicate balance between salt accumulation and salt removal by the processes of soil formation, in particular, leaching. When salt accession exceeds leaching, salt accumulates more rapidly than it can be removed, and saline soils are the result. When the accumulation is so great that other profile features are hidden, the soil is called a *solonchak* (which is a saline soil). But when salt accession is fairly strong, and the soil-forming processes have some effect on accumulation (such as illuviating the free salts to the lower profile), and the other processes of soil formation are not hidden, the soil is then called a *solonetz.* A stage further in the balance between soil formation and salt accumulation leads to the *solodised solonetz*, in which the salt is redistributed through the profile, and other soil-forming processes leave marked effects on the profile. The final stage leads to *solodic soils* in which most of the salt is removed by leaching and the profile is strongly differentiated by the other soil-forming processes. Both solonetz and solodised solonetz may be called alkaline soils, but solodic soils are very like podzolic soils in their properties.

SOLONCHAKS

In America, solonchaks are called white alkali soils because excess free sodium salts colour them white, especially on the surface, where crusts usually form. These soils attain their maximum development in periodically inundated coastal areas with low rainfalls such as the lowlands around the Gulf of Carpentaria, and in desert salt pans or playas which are common in South Australia. In these instances, salt accession far outweighs salt removal by leaching and there is little profile development aside from a marked concentration of salt at or near the surface, a consequence of evaporation. The crust is usually quite shiny from the sodium (or halite) crystals. A thin horizon of sodium-saturated clay is frequently found below the crust. This horizon may have a puffy character from the dispersion caused by sodium-saturation; but if excess free sodium salts are present, loosely-packed peds may occur. Solonchaks associated with coastal inundated areas and playas as described above are termed *primary solonchaks*, because in their profiles there is little evidence of any soil-forming processes, other than *salinisation* (the term used for the accession of salt and its accumulation within the profile). Primary solonchaks consist essentially of a salt crust overlying a thin soil horizon resting on the parent material, which also is usually salty or calcareous or both.

Salinisation may cause any soil profile to degenerate into a solonchak. Salinisation is essentially the domination of salt accumulation over leaching and soil formation, either temporarily or for a prolonged period. Natural changes in drainage or climate, and man's poor management of the land may induce salinisation in any type of soil profile, causing degeneration into a solonchak. In such cases, some of the original profile characteristics usually remain beneath the salt accumulation and the soil is called a *secondary solonchak*, a polygenetic soil. In Western Australia, the clearing of the native vegetation from the hills of dry areas, together with subsequent cultivation, has initiated a redistribution of soluble salts along catenas, resulting in increased deposition on the lower slopes and alluvial flats, to form solonchaks. This has made the lower-slope lands unfit for cultivation, and they will remain so until the excess salts have been removed; that is, until man, who caused the salinisation, de-salinises the soil. Unwise irrigation has formed solonchaks in many depressions in the Riverina where the water table has been raised to a dangerous level and drainage is poor. This applies especially to the grey and brown soils of heavy texture which have a high capillary fringe: if this fringe is raised to the surface, mass evaporation of the salty ground-water or irrigation-water leaves salt deposits at or near the soil surface.

Both primary and secondary solonchaks are saline soils and are distinguished in the field by their salt crusts, their close to neutral *pH*, and their absence of vegetation (excepting the most salt-tolerant varieties such as salt bush). Saline soils have been successfully reclaimed for cultivation. The reclamation procedure consists essentially of man-induced solonisation which turns the solonchaks into solonetz or even solodised solonetz. The excess free sodium salts first have to be leached by repeated irrigations. This leaves a very poorly structured sodium-dominated, high *pH* soil, still unsuitable for cultivation. The next step is to replace the sodium on the soil colloids by a cation which is not toxic to plant growth and which promotes better structure development. The obvious choice is calcium: this cation is available in commercial fertilizers as lime ($CaCO_3$) and gypsum ($CaSO_4 \cdot 2H_2O$). If the farmer adds lime to his soil to replace the sodium, the results are most unsuccessful: the calcium from the lime replaces some of the sodium on the colloids, and sodium carbonate forms in the soil, as shown by the equation:

$$2\text{Na–clay} + CaCO_3 \longrightarrow \text{Ca–clay} + Na_2CO_3$$

But sodium carbonate, being a basic salt (the salt of a weak acid and strong base), has a *pH* of 10 and lime is only very sparingly soluble at such a high *pH* value; therefore, the replacement reaction slows down to a negligible rate, and the soil is left in a worse state than before the lime was added. Its *pH* being then 10 rather than about 9, cultivation at such high *pH* values would be unsuccessful: sodium would still dominate the colloid, so there would be little improvement in soil structure.

Reclamation by the addition of gypsum, however, is generally successful, for the solubility of gypsum is less sensitive to *pH* than is that of lime. When gypsum is added, the sodium on the clay is still replaced by calcium, but this time sodium sulphate, a neutral salt (the salt of a strong acid and a strong base), forms in the soil as shown in the simplified equation:

$$2\text{Na–clay} + CaSO_4 \longrightarrow \text{Ca–clay} + Na_2SO_4$$

The sodium sulphate maintains the soil *pH* close to 7 until the replacement reaction has proceeded sufficiently for calcium to dominate the colloid, with the related marked structure improvement. The final step is to leach the soluble sodium sulphate from the soil by repeated irrigations. This procedure has been successfully used on saline soils in the Riverina, especially on those damaged by over-irrigation during the early history of rice cultivation. It results in a well-flocculated soil with *pH* close to 8, though the general lack of organic matter means that natural reserves of nitrogen and phosphorus are low.

Solonetz

Solonetz soils also contain enough sodium to interfere with plant growth, but the sodium is mainly adsorbed onto the soil colloids rather than existing as free salts (cf. solonchaks). Solonetz, although not as chemically poor as solonchaks, are poorer physically because they are dispersed rather than flocculated. Dispersion induces a puffy character which manifests itself in the round-topped domed columnar structure so characteristic of solonetz. The cracks between these large columns (some columns are 18 inches across) provide the only important natural planes in these soils, so water penetration is usually quite difficult. The columns and the secondary blocky structure within the columns are quite unstable when removed from the soil profile, so they do not constitute peds in the true sense of stable soil aggregates. The sides of the columns in solonetz are often stained black by illuviated humus, so in America these soils are termed black alkali soils.

In Fig. 11.4, a typical solonetz from Saskatchewan is described. It has a high *pH* in the *B* horizons, indicating sodium domination of the colloids; and the lower *pH* in the eluvial horizons,

Location: 1 mile northeast of Lawson, Saskatchewan, Canada. **Land-use:** wheat rotation cropping.
Parent material: calcareous glacial till. **Slope:** flat.
Drainage: well-drained. **Average Annual Rainfall:** 14.5".

Horizon	Depth	Colour	Texture	Structure	Lime	pH
A	0"-7"	brown	loam	apedal massive	nil	7.5
E (horizon is sporadic)	less than 1"	grey	loam	apedal massive	nil	7.5
Bh, t	7"-12"	dark (coffee) brown	light clay	strong, round-topped columnar, breaking into angular blocky	trace of fine earth	8.5
Bca, cs	12"-35"	yellowish-brown mottled with red and white	clay loam	weak angular blocky	lime and gypsum throughout	9.0
C	35"+	grey, calcareous, loamy, glacial till				

Figure 11.4: Field profile description of a solonetz soil from Saskatchewan, Canada (after Riecken and Stalwick, 1945).

suggesting sodium has been mostly eluviated. Lime and gypsum have been illuviated to the *B* horizons where they are deposited together rather than in the usual order of lime above gypsum. A fairly thin lighter textured *A* horizon has developed, but leaching is insufficient to form a proper *E* horizon. The *E* horizon occurs only sporadically, at favourable locations, and is always less than one inch thick. The *Bh, t* horizon is coloured by the illuviated humus and contains the characteristic domed columns. It contains considerably more clay than the *A* horizon, suggesting clay illuviation. The *Bca, cs* horizon is also higher in clay than the *A* and *C* horizons as the following analyses indicate:

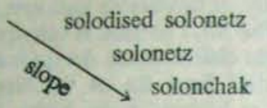
Horizon	*Percentage clay*
A	12.7
Bh, t	33.6
Bca, cs	23.3
C	19.8

This Canadian example was selected to illustrate the solonetz profile because true solonetz with neutral eluvial horizons and alkaline subsoils are rare in Australia (partly solodised solonetz with weakly acid eluvial horizons are more common). As in the red-brown earths, some Australian pedologists discount the importance of clay illuviation in the formation of solonetz and, instead, stress differential weathering with some destruction of clay in the eluvial horizons, and layering of the parent materials. However, some illuviation of clay probably does occur.

In solonetz, then, profile differentiation is more advanced than in solonchaks. Not only are the salts redistributed through the profile and some of them removed completely, but clay, humus, lime and gypsum are illuviated. However, eluviation of constituents from the upper horizons is not intense enough for an *E* horizon to develop (cf. solodised solonetz). The general term for these processes which redistribute the salts through the profile to form a solonetz is *solonisation*; the result of solonisation is an alkaline (sodic, or non-saline alkali) soil.

Solonetz may develop by solonisation of fresh parent material such as alluvium, loess or weathering bedrock; such solonetz have relatively simple profiles like the one described in Fig. 11.4. However, any type of soil can develop solonetz characteristics by solonisation, that is, by salt accession and its redistribution through the profile. Such solonetz have more complicated profiles, distinguished by relict properties of the former soil; they are therefore polygenetic soils. An increase in leaching (natural or man-induced) can change a solonchak into a solonetz (see Chapter 12 where this is discussed in relation to Downes' theory).

Solonetz are found in conditions more favourable to leaching than solonchaks. They are therefore likely to be found in wetter areas or in better-drained positions than solonchaks with which they may be associated in a catena. The soils discussed in this section may occur in catenary succession as follows:



In very dry areas, solonchaks may form the lowest member of the catena, but in wetter areas, solonetz are found in this position, with solodised solonetz on the better-drained slopes.

Solonetz soils, in their natural state, are used for grazing, but their carrying capacity is low because of their sparse, salt-tolerant vegetation. Like solonchaks, they are, usually, first reclaimed (for example, by irrigation and the addition of gypsum) before they can be successfully cultivated. The addition of sulphur or sulphuric acid has also been tried successfully to replace the dominating sodium, and lower the *pH* of these soils.

SOLODISED SOLONETZ AND SOLODICS

Stronger leaching, relative to salt accumulation, not only removes the excess free salts from the profile, but replaces some of the bases (especially sodium) attached to the soil colloids by hydrogen, at the same time lowering the *pH* of the soil, especially in the eluvial horizons where replacement is greatest. Concomitant with this salt removal, a profile differentiation takes place which is stronger than in solonetz: an *E* horizon develops, lime and gypsum are illuviated more deeply and there is a strong texture contrast between the eluvial and *B* horizons. The processes involved are called *solodisation*: these processes are capable of changing an alkaline solonetz into an acid podzolic soil. At the stage of solodisation when only some of the bases on the colloid are replaced by hydrogen, the soil is called a *solodised solonetz*. When most of the bases have been replaced by hydrogen, and the soil is more like a podzolic than a solonetz, the soil is called a *solodic* (soloth, solod or soloti)[3]. Solodics usually have a lower

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[3] These terms are used differently in Stace, Hubble and others' *Handbook of Australian Soils* (1968). They include solodics with solodised solonetz, and claim that the only difference lies in the structure of the *B* horizon (columnar in the former and blocky in the latter). They use the term soloths or solods for the more leached soils, here called solodics. This writer contends that to distinguish between solod and solodic is confusing and prefers the more traditional use of these terms (e.g., Stephens').

Location: 36 miles from Coonamble along the road to Pilliga.

Vegetation: bimble box and pine woodland.

Parent Material and Underlying Rock: Jurassic shale, sandstone and conglomerate.

Slope: flat.

Drainage: well drained.

Average Annual Rainfall: 18".

Horizon	Depth	Colour	Texture	Structure	Lime	pH
A	0"-5"	brown	sandy loam	apedal massive	nil	6.5
AE	5"-13"	brown to light brown	sandy loam	apedal massive	nil	6.0
E	13"-16"	light grey	sand	apedal single grain	nil	6.5
Bt	16"-32"	brown to reddish-brown	light clay	round-topped columnar, breaking into irregular angular blocky	nil	6.5
Bg, 1	32"-42"	yellowish-brown	light clay	weak irregular angular blocky	trace of fine earth	8.0
C	42"+	weathering rock				

Figure 11.5: Field profile description of a solodised solonetz soil near Coonamble, New South Wales (after Downes and Sleeman, 1955).

pH, better-developed *E* horizon, sharper boundary between the *E* and *B* horizons, deeper weathering, and less lime in the lower part of the profile than solodised solonetz, although all these differences are not always present when comparing any two profiles.

In Fig. 11.5 a solodised solonetz from the western slopes of Central New South Wales is described. There is a much lower base content compared with the solonetz described in Fig. 11.4 which has higher *pH* values, and the free lime and gypsum are much higher in the profile. Gypsum is absent from the solodised solonetz and very little lime is found in the *Bg, t* horizon (not even enough to raise the *pH* to 8.5, the usual *pH* associated with free lime). The following analyses indicate the strong texture contrast between the horizons in the soil described in Fig. 11.5:

Horizon	*Percentage clay*
A	14
AE	12
E	8
Bt	33
Bg, t	37

There are three eluvial horizons in this profile: a transitional *AE* horizon, more like the *A* than the *E* horizon, lies between the *A* and bleached *E* horizons.

The processes forming solodised solonetz are similar to those forming solonetz, but their intensity is greater and there is considerable replacement of bases by hydrogen on the soil colloids. Thus, illuviation of lime and clay are marked, gypsum is completely removed from the profile and there is some evidence of humus illuviation (overseas solodised solonetz usually have more marked features relating to humus illuviation than this profile). Eluviation, especially of clay, in the *AE* and *E* horizons is quite strong. Formerly, in Australia, solodised solonetz were regarded as polygenetic soils, but it is now recognised that they can form monogenetically from fresh parent material. The profile described in Fig. 11.5 is probably an example. The main requirement is that the balance between salt accession and salt removal is such that free salts are leached from the profile, but the colloids, especially in the lower horizons, still contain appreciable bases, and lime may be deposited in the lower profile. Although solonetz profiles provide a suitable starting point for solodisation, similar results can occur without the soil ever going through the solonetz stage. However, many Australian solodised solonetz are still believed to be polygenetic soils that have been salinised and de-salinised. These soils and their associated environmental changes are discussed in more detail in Chapter 12 (relict soils).

Location: 1 mile northeast of Glenlee about 40 miles southwest of Sydney.	**Vegetation and Land-use:** sclerophyll woodland: improved pastures grazed by cattle.
Parent Material and Underlying Rock: greywacke.	**Slope:** flat.
Drainage: fairly poor.	**Average Annual Rainfall:** 27″.

Horizon	Depth	Colour	Texture	Structure	Lime	pH
A	0″-3″	mid brown mottled with dark grey	fine sandy loam	apedal massive	nil	5.0
E	3″-6″	light brown and light grey	fine sandy loam	apedal massive	nil	5.0
Bt	6″-8″	dark brown mottled with yellow and grey	heavy clay	strong 8″-9″ diameter round-topped columnar breaking into weak 1″ angular blocky, decreasing with depth	nil	5.5
Bg, t	8″-39″	yellowish-grey mottled with red and brown	medium clay		a few scattered concretions	6.0
BC	39″56″	yellowish-brown mottled with grey, red and black	light clay	weak angular blocky	a few scattered concretions	6.0
C	56″+	weathering greywacke				

Figure 11.6: Field profile description of a solodic soil southwest of Sydney, New South Wales (after Walker, 1961).

Solodised solonetz require better leaching conditions than solonetz. Such conditions can arise from higher rainfall, or better drainage, or both. Solodised solonetz are widespread in Australia between the 15″ and 35″ average annual isohyets. Their most distinguishing features are the domed columnar structure in the upper *B* horizon, weakly acid *pH* in the upper profile and weakly alkaline *pH* in the lower profile, and absence of lime excepting in the lowest horizons.

Solodics, requiring even stronger leaching than solodised solonetz, are found in the zone of podzolised soils. Their previous history of salinisation and de-salinisation was advanced as a possible explanation for their greater depth of weathering (they are usually deeper than neighbouring podzolics, even in drier areas), their peculiar domed columnar structure, and their more-than-usual amounts of bases on the colloids of the lower horizons. They are always considered to be polygenetic soils, but are treated here (as well as in Chapter 12) to complete the succession of processes involving salty soils.

Solodics are more completely solodised than solodised solonetz; that is, the processes of solodisation are greatest in solodics. A solodic (not a very typical one) from the Sydney area is described in Fig. 11.6. There is an acid *pH* throughout the profile, but the *pH* is not quite as low as that of neighbouring podzolics. The domed columnar structure of this solodic is well-preserved; this is not always the case, for this structure often degenerates to blocky structure associated with hydrogen-dominated colloids. Although some lime concretions are found in the lower horizons of this profile, they have little effect on raising the *pH* of the soil, and the soil surrounding the nodules is apparently well leached. Clay illuviation is quite strong in this profile, but (from analyses) organic matter is greatest near the surface and decreases very much with depth. However, an additional process, not usually found in solodised solonetz, occurs in this soil, namely, illuviation of iron (from analyses, the *B* horizons contain twice as much iron as the *A* horizons). This profile is not well-drained, so the *B* horizons are mottled with drab greys and yellows. The *A* and *E* horizons in this profile are not as well-developed as those of solodics found on the western slopes of New South Wales, where each horizon may be up to eight inches deep.

The delicate balance between salt accumulation and salt removal which gives rise to the four soils (solonchak, solonetz, solodised solonetz, and solodics) discussed in this section is illustrated in Fig. 11.7. The horizon complexity clearly increases as leaching increases and salt accumulation decreases. This diagram is not intended to be viewed solely as the stages in the genesis of

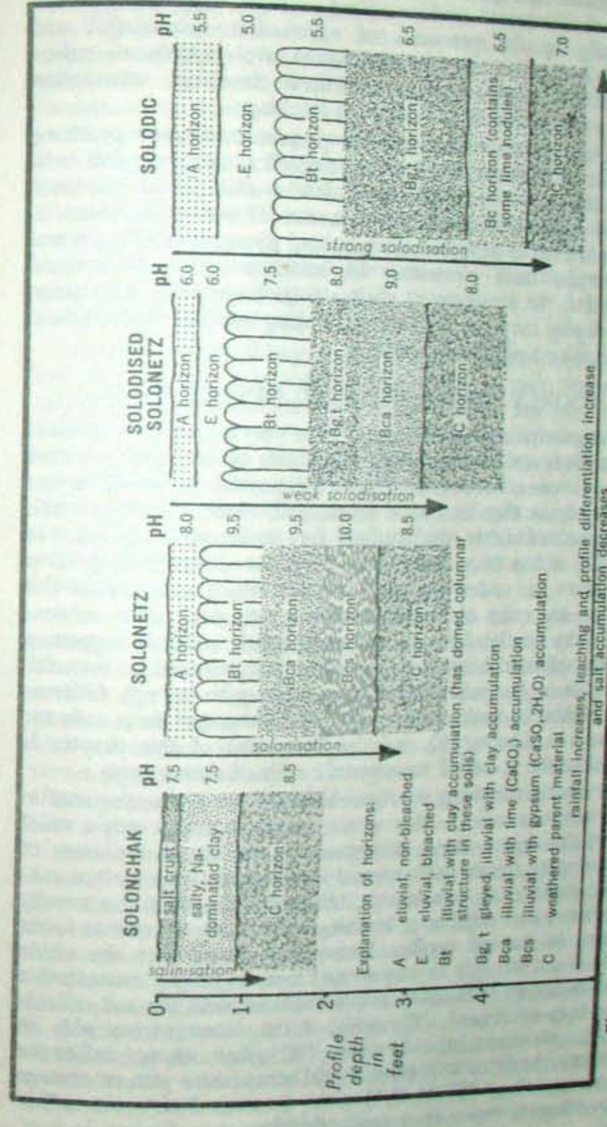

Figure 11.7: The formation of solonchaks, solonetz, solodised solonetz and solodics by the processes of salinisation, solonisation and solodisation.

solodic soils, though this is part of the interpretation intended. Solonchaks, solonetz and solodised solonetz may form monogenetically by the processes of salinisation, solonisation and solodisation respectively, but solodisation involves both salinisation and solonisation as well as additional leaching; solonisation involves salinisation with additional leaching.

Solodised solonetz and solodics present cultivation problems similar to those encountered on the podzolics, excepting that their *pH* is higher; so the problem of low availability of nutrients, caused by strong acidity, is not as acute. However, in Australia, these soils tend to have low nitrogen and phosphorus reserves and lack certain trace elements. In addition to these chemical deficiencies, the structure of their eluvial horizons is very poor, so on drying they set quite hard, making moisture penetration, when it rains again, more difficult.

POORLY DIFFERENTIATED DESERT SOILS

Monogenetic desert soils tend to have very poorly differentiated profiles. It is not surprising that these soils are less common than polygenetic ones, because erosion and deposition are much slower in desert areas than in humid areas, and, moreover, desert soils have remained fairly undisturbed for thousands of years. In Australia, it has been found that very few desert soils develop *in situ* from the underlying rock. Profiles which approximate this condition are quite often strewn with quartz and other resistant stones quite unrelated to the parent material, and this suggests a fairly complicated history of deposition and removal of material. Many soils of Australian desert areas have quite strongly differentiated profiles because of their polygenetic history: these soils are discussed in Chapter 12 and the remainder of this chapter is devoted to the so-called monogenetic soils of desert areas.

The conditions giving rise to solonchaks are not widespread in desert environments; in other words, playas represent only a small area of the desert. More widespread in the desert are areas of gently sloping terrain with internal drainage, or little surface runoff. In these areas, rain-storms, although infrequent, are usually fairly heavy and capable of leaching soluble constituents to lower horizons in the soil profile. Removal of bases from the whole profile is not strongly developed and there is usually considerable accumulation of salts in the lower horizons, with the soil colloids being base-dominated. The soils of the desert (areas with an average annual rainfall of less than 10″), then, are not unlike the solonetz and related soils of semi-arid areas (areas with an average annual rainfall of 10″ to 20″) partly because desert rains fall in heavy showers rather than gentle drizzles.

The remaining monogenetic desert soils may be divided into two groups according to texture: desert loams which contain appreciable quantities of silt and clay, and desert sands which contain very little, if any, silt and clay. In hot deserts, both of these soils are usually red in colour throughout their profiles. Because of the heat and aridity, iron is oxidised very soon after it is released by weathering from aluminosilicate minerals; there is thus little opportunity for the iron to move through the soil profile (cf. podzolised and lateritic soils), and there is little water available to hydrate the iron oxides to form the yellowish-coloured oxides. Since organic matter is lacking, soil colour, then, is dominated by the less hydrated iron oxides. The actual percentage of these oxides may be quite small, especially in the desert sands, though an intense red colour is produced.

In *desert loams*, if any texture contrast occurs between the *A* and *B* horizons, it is usually very slight as clay illuviation is relatively unimportant in these soils. The *A* horizon of desert loams is usually loamy to clayey, and granular or blocky in structure. Its *pH* is usually close to or just above neutral. The upper *B* horizon may resemble that of solonetz, that is, it may have a weak domed columnar structure and quite high *pH*. The lower *B* horizons (*Bca* and *Bcs*) contain accumulations of lime and gypsum and sometimes other free salts such as sodium chloride.

Desert loams in Australia are found on fine textured parent materials in areas with an average annual rainfall of 5″ to 10″. They are common on alluvium and fine-grained sedimentary rocks such as shales. The dominating soil-forming processes are the illuviation of bases, with the associated effects on soil structure. The soil surface is frequently covered by a mantle of resistant stones (quartz especially) which remain as desert pavement (reg) after some of the surface soil has been deflated or water-eroded after heavy rains.

About one-third of Australia's desert area is sand-covered. Some of this sand forms active dunes; however, many of the dunes have been stabilised by vegetation, and even in the desert, when this occurs, soil profile formation follows. The dunes are mostly siliceous sands, the red colour coming from very thin iron coatings on the quartz grains. Because leaching is so ineffective in these areas, profile differentiation is not very strong; lime, for example, is usually illuviated to a few inches or a foot below the surface where the profile tends to become more compact (i.e., apedal massive rather than apedal single grain). The high permeability of the sands facilitates base removal, so the *pH* is often lower than that of desert loams, and free sodium salts are illuviated to quite low depths in the profile, and some are completely removed. These soils, which contain very little profile development, may be called *desert sands*.

The soil-forming processes in desert sand dunes contrast very strongly with the soil-forming processes in coastal sand dunes in humid areas. In coastal sand dunes, horizon differentiation is very marked for a depth of several feet, and even the resistant constituents, such as iron oxides, are illuviated leaving a deep *E* horizon in the podzol profile (see Chapter 8). Such differentiation is not possible under the low rainfalls of desert areas, though the parent material offers similar susceptibility to leaching.

Desert sand profiles are found in the main dune areas of the Australian desert, such as the Simpson Desert south of the Macdonnell Ranges and in northern Western Australia. Like the desert loams, their land-use is limited to grazing, with a very low carrying capacity because of the scanty vegetation and lack of water. If over-grazed, both of these soils are susceptible to damage by wind and water erosion.

Desert loams and desert sands are both red soils found mainly in hot deserts such as the Australian desert. In cooler deserts, such as the Central Asian, Patagonian, and northern U.S.-Canadian deserts, the lower temperatures inhibit the rapid oxidation of iron and consequent formation of red iron oxides, so some of the soils lack the prominent red colours found in soils of hot deserts. Aside from this difference, the soils are similar: there is some base removal and distribution through the profile and sometimes weak clay illuviation. Soil colour is dominated by the aluminosilicate minerals and salts, rather than by the iron oxides, and is therefore grey or white rather than red. If humus is present (the cooler conditions lower its rate of destruction), the soils may be brownish in the surface horizons, becoming grey or white with depth. Such soils are called *sierozems* or *grey desert soils*. They are poorly represented in Australia because there are no high-latitude deserts: profiles similar to American sierozems are found on iron-deficient parent materials within areas of desert loams and desert sands, but there is no zonal area for them (see Chapter 5, Fig. 5.1).

Desert loams and desert sands, the zonal soils of the Australian desert, account for only a small part of desert soil variation. Many of the Australian desert soils owe their character to relict, deeply-weathered profiles, now partly destroyed. Because of the low weathering intensity in desert areas, relict soils are more likely to be preserved there than in humid regions where most of them have been destroyed. These relict desert soils are discussed in the next chapter.

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Chapter Twelve

Relict Soils in Australia

CONTENTS

IN CHAPTER 5 IT WAS POINTED OUT THAT NOT ALL SOILS DEVELOP from the first stages of soil formation to the mature profile, under a constant environment. Climate in the immediate past (right back to the Tertiary) has fluctuated markedly, and erosion and deposition have been active in response to these fluctuations and to changes in base level. Some of Australia's previous climates were very different from the present climates, especially during the Tertiary and Pleistocene when parts of Australia were hotter and wetter, and cooler and wetter respectively, than anywhere in Australia to-day. It is therefore not surprising that some of Australia's very old soils differ markedly in morphology from their younger neighbours. Soils whose morphologies are not entirely explicable in terms of their present environment are called relict soils, fossil soils or palaeosols. The term relict soils is used in this book.

Every continent contains relict soils which reflect its climatic and geomorphic history. In Northern Europe and North America, the most widely studied relict soils are related to the Pleistocene glaciations: these provided morainic and loessial materials which developed soil profiles in stable phases within and following the Pleistocene. The glaciations destroyed the pre-Pleistocene land-surfaces and soils. In Australia, on the other hand, the effects of the Pleistocene glaciations are limited to western and central Tasmania and the Monaro region; elsewhere, the Pleistocene glaicial phases were characterised by increased rainfall (they are referred to as the 'Pleistocene pluvials'), and the interglacials by warmer, drier conditions.

Since most of Australia was unaffected by ice during the Pleistocene, soils relict of the Tertiary are more widespread than in Northern Europe or North America. For reasons of space, the following discussion of relict soils is restricted to examples

from Australia; quite a different pattern would emerge if the soils of a glaciated continent were examined. There is not room, either, to treat tephro-chronology, which is the study of soil successions on volcanic ash deposits; for obvious reasons, there is little scope for this study in Australia, but New Zealand offers rich opportunities.

Relict soils cannot reflect every fluctuation in past climate, especially those which brought about less intense weathering and profile differentiation, because subsequent processes have destroyed the imprints of earlier environment. One would therefore expect relict soils, resulting from weathering more intense than the present, to be more common than relict soils resulting from weathering less intense than the present. Thus soil evidence of previous drier climates is largely obliterated by leaching in subsequent wetter climates, and the older the soil the greater the opportunity for modification by the later environments.

In Australia, past climates are most clearly expressed in the relict soil profiles of the Tertiary sub-tropical climate, the Pleistocene pluvial and arid phases, and the Recent arid climate. The most spectacular relict soil profiles were formed during the Tertiary, when the intensity and depth of weathering was greater than in any climate since. Tertiary relict soils are best preserved in the arid parts of Australia, where the rate of erosion of the old profiles is very much slower than in humid areas. As there is now copious literature on Tertiary relict soils in Australia, the examination of these soils in this chapter can proceed by a study of some specific examples rather than through a generalised review. Students interested in a more comprehensive coverage should consult the references at the end of the chapter.

To illustrate the effects of climatic fluctuation on soil morphology during and after the Pleistocene, the soils of the Riverine Plain may be studied. This area has received considerable attention since 1950 and many detailed studies in soil stratigraphy are available. The soil succession has developed on alluvial and aeolian parent material. Similar periodic successions of soils are found all along the western slopes and plains of eastern Australia, but, with the exception of the Riverine Plain, these have received little attention. We shall limit our examples to the Riverine Plain.

There is a good deal of soil evidence of an arid period in the Recent. This is suggested mainly by the solodic soils, which were considered in Chapter 11. Here, we shall discuss in more detail the processes of salinisation and desalinisation and relate them to climatic change.

This chapter is divided into three sections: Tertiary relict soils, soils of the Riverine Plain, and soils of the Recent arid period. Each of these demonstrates the effect that a previous environment has on soil formation. Most of the publications referred to for these three examples of relict soils have appeared in the past ten years, though the theoretical concepts are very much older, as was indicated in Chapter 5. Since a great deal of work is in progress in the three fields, especially in the field of Tertiary relict soils, the ideas presented here are only tentative.

TERTIARY RELICT SOILS

It is well recognised that Australia, during Tertiary times, experienced a much hotter and wetter climate than its present climate. The mainland probably had tropical and sub-tropical climates, rather like the equatorial and savanna or monsoon climates of the present inter-tropical areas. Relief over the whole continent was much more subdued than it is now, for the Eastern Highlands had not then been elevated. These conditions of relief and climate did not prevail for the first time during the Tertiary: geological evidence suggests that before the Tertiary there were several episodes of hot, wet conditions on the Australian mainland. However, most soils relating to these earlier episodes have been obliterated by subsequent erosion, with the exception perhaps of some of the Mesozoic soils, remnants of which are found in parts of the Australian desert. Elsewhere, the pre-Tertiary erosion surfaces have been denuded and Tertiary relict soils are the oldest soils still in existence. Deeply weathered soils outside tropical Australia, and some of the soils within tropical Australia, are generally ascribed to the Tertiary and usually form the oldest landsurface remaining.

It seems that for most of the Tertiary the climate was both hotter and wetter than the present climate, although few workers agree on which part of the Tertiary was hottest and wettest and therefore had the greatest intensity of weathering. There is considerable dispute over the age of the deeply weathered profiles. Early writers (e.g., Woolnough, 1927) suggested a Miocene age for the deeply weathered profiles and their landsurfaces, but later workers favour an earlier date. Dury, Langford-Smith and McDougall (1967) suggest that the deeply weathered soils are at least as old as Oligocene and perhaps considerably older: their chronology is supported by potassium-argon datings. It is quite possible that the deeply weathered profiles could differ in age, if suitable conditions for their formation prevailed through the whole of the Tertiary.

Although suitable conditions possibly prevailed for a great length of time during the Tertiary, so far only one deeply weathered landsurface has been identified with certainty in Australia, suggesting only one major episode of deep weathering, an episode which probably occurred over a very long time. The deeply weathered profile, which in parts of central Australia attains 200 feet, is capped by an indurated horizon, called *duricrust* (from the Latin, *duro*, to make hard or harden). The composition of this duricrust varies across Australia: it is lateritic (rich in hydrated iron oxides) and bauxitic (rich in hydrated aluminium oxides) in the wetter, coastal areas, and siliceous (rich in silica) in the drier, inland areas.

The term duricrust is generally applied only to the hard cappings which overlie the deeply weathered soil profiles. Indurated horizons in the regolith are not always associated with deeply weathered soil profiles: these horizons are usually not called duricrust but are named according to their composition. *Silcrete* is an horizon hardened by silica enrichment, whether or not it is associated with a deeply weathered soil profile: some silcretes are in fact duricrusts, whereas others are not. Similarly, *ferricrete* is an iron-enriched, indurated horizon (*laterite* is the term for ferricretes associated with deeply weathered soil profiles, i.e., for ferricretes that are duricrusts). *Calcrete* is a calcium carbonate-enriched indurated horizon: calcrete is not usually associated with deeply weathered soil profiles.

A great deal of confusion has arisen through relating silcretes and ferricretes to 'periods of laterisation' without regard for their association with deeply weathered soil profiles. For this reason, many workers claim there were several 'periods of laterisation' resulting in several Tertiary landsurfaces (e.g., Hallsworth and Costin, 1953). In all these cases, indurated horizons alone provide the evidence, and there appears to be only one deeply weathered landsurface capped by duricrust.

Ferricretes and silcretes, very similar in properties to those capping deeply weathered soil profiles (i.e., actual duricrusts), may form in many ways which are unrelated to deep weathering. This was especially so during the Tertiary when high weathering intensity caused a great deal of mobility of iron and silicon which, under low weathering intensities, are fairly immobile. In the Inverell area of the northwestern slopes of New South Wales two 'periods of laterisation' have been claimed, largely on the basis of ferricretes between basalt flows. Workers in the area (e.g., Hallsworth and Costin, 1953) claimed that the earlier basalts were laterised, and then newer basalts flowed over the lateritic profiles, and some of these in turn were also laterised.

Close examination of the area revealed only one deeply weathered profile, and the so-called laterites between the basalts were *bole*; bole is the soil on the underlying flow baked and enriched with iron by the overlying flow.

These bole horizons are not restricted to basalt: soils on granite develop similar characteristics when basalt flows over them. McGarity and Munns (1959) recognised bole in basalts in the Lismore area and noted that it weathered to a krasnozem soil even though the rainfall is less than 55″ per annum (which is the amount needed to form krasnozems from basalt). These krasnozems are therefore not relict soils, for they are products of the current cycle of weathering, even though their parent material was once a soil before being 'metamorphosed' by the overlying flow to form bole. McGarity and Munns used the term 'Minyon bole' for the material between the flows, but the term bole is sufficiently well established in geological literature to stand on its own.

Boles are very common between basalt flows in the Inverell area. They are rarely more than 8-10 feet thick, but as they outcrop on hillslopes, lateral movement makes them appear thicker than they really are. Fig. 12.1 shows two bole outcrops along a hillslope of three flows, north of Inverell. Note how the boulders of bole have rolled downslope. The soil downslope from the bole is redder than soil upslope from the bole, and contains weathered ironstone in the upper horizons.

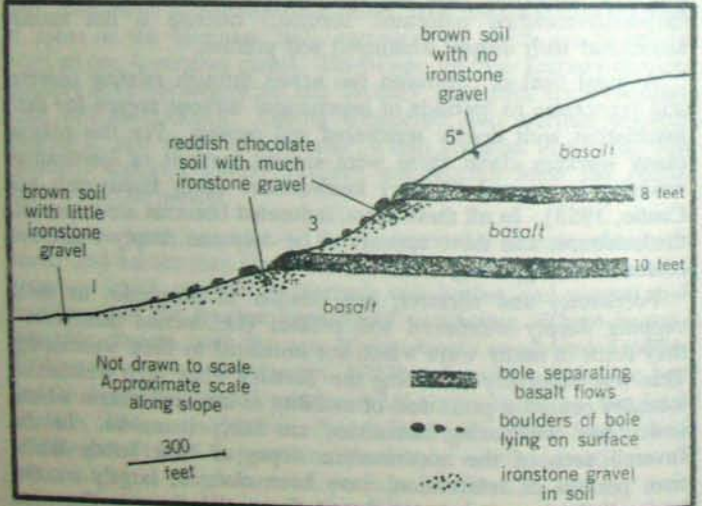

Figure 12.1: Bole horizons separating basalt flows along a hillslope north of Inverell, New South Wales.

The confusion between laterite and bole is increased by their similarity in the hand specimen: both may be massive, vesicular or nodular; both are reddish in colour; and both when fresh are so hard that a hammer is needed to break them. They both weather to softer, yellowish oxides. They are both accumulations of hydrated iron oxides, laterite by subaerial enrichment, and bole by geological enrichment. Some of the iron oxides in bole result from concentration in the soil before the upper flow covered it, especially if the soil formed during tropical or sub-tropical weathering conditions. Much caution is therefore needed when identifying deeply weathered profiles in areas of basalt flows: the presence of ferricrete is not sufficient evidence, and deep augering is needed to determine whether the ferricrete overlies fresh basalt (and is therefore bole) or deep mottled and pallid horizons (and is therefore laterite).

Ferricrete can form in ways other than by deep weathering or geological enrichment. Milne (see Fig. 5.7) showed that seepage water from red earth profiles formed deposits of hydrated iron oxides at break-of-slope in catenas in East Africa, without either deep weathering or geological enrichment. Unlike deep weathering which takes thousands of years, this type of subaerial accumulation of ferricrete may take only a very short time. In Australia, accumulations of iron oxides resembling laterite in structure and hardness have formed in less than 10 years in seepage zones leading from dams and culverts.

Similar confusion has arisen in the interpretation of silcretes. Silification was apparently widespread during the Tertiary and affected rocks and weathered materials in all parts of the landscape without always being associated with deep weathering. Silcrete horizons in arid Australia are just as common as ferricrete horizons are in humid Australia. Some pedologists even claim that silification is still active under certain circumstances in the Australian desert.

It is therefore very important when identifying Tertiary relict soils to note not only the indurated horizons, but the whole sequence of horizons which make up the deeply weathered profile, of which the duricrust is only the capping. Although the composition of the duricrust varies across Australia, the sequence of horizons in the deeply weathered profiles is essentially the same: the duricrust overlies mottled horizons which rest on the pallid horizons of weathered parent material, as described in Chapter 7.

The reasons for the change in the composition of the duricrust across Australia are not clearly understood. One possible explanation is that during the Tertiary, although Australia's climate was tropical or sub-tropical, it was by no means uniform across the continent. Rainfall and temperature could have decreased in a manner similar to the present decrease across the continent, although the overall amounts would have been higher. Thus the weathering intensity could have been greater in northern and eastern Australia, causing iron and aluminium to be mobilised and concentrated in the duricrust while much of the silica was lost to drainage. In the cooler, drier areas, leaching would not have been as intense, so iron and aluminium would scarcely have been mobilised, and silica (instead of being removed from the profile by leaching) would have concentrated in the duricrust, in a manner similar to the concentration of iron and aluminium in laterite and bauxite respectively.

Pedologists generally agree that more intense leaching is required to mobilise iron and aluminium than is required to mobilise silica. This theory, however, is not corroborated by the depth of weathering of the relict soil profiles: silcreted profiles in arid Australia tend, on the whole, to be deeper than lateritic and bauxitic profiles, but perhaps the faster denudation rate in wetter areas prevented deep profiles from forming.

In the following discussion, Tertiary relict soils are separated into those of humid areas (lateritic and bauxitic profiles) and those of arid areas (silcrete profiles) because not only did the Tertiary environments of these soils differ, but the profiles have undergone very different post-Tertiary modifications. The boundary between the two types of deep weathering profiles has not yet been delimited. Its delimitation will probably prove difficult because the profiles frequently grade into each other and because both types of profile are found in some areas.

TERTIARY RELICT SOILS IN HUMID AREAS (LATERITIC AND BAUXITIC PROFILES)

Relict soils in the wetter parts of Australia include not only the deeply weathered profiles with marked horizon differentiation formed under a fluctuating water table (the lateritic and bauxitic profiles) but also the deeply weathered, little-differentiated profiles formed under free drainage (krasnozems and red earths). The latter soils may not contain duricrust, but are closely related to the duricrusted profiles. The processes of laterisation and ferralisation which form these soils were discussed in Chapter 7; our chief concern here is with the role these relict soils play in present soil distribution in Australia.

Laterisation is not regarded as an active process in Australian soils under the present climates, but ferralisation is active in areas where the average annual rainfall is more than 55″, especially where there are silica-poor parent materials producing clayey soils. However, during the Tertiary, most of the moister parts of Australia were probably mantled with both types of soils. Uplift and dissection of the Tertiary landsurface has removed these soils from many areas, so they tend to remain only on undissected interfluves. Some of the ferruginous material eroded from these soils has been deposited as alluvium or colluvium, and it still has an important effect on the distribution of the present soils. The iron oxide nodules are particularly resistant to destruction and may undergo several phases of erosion, transport and deposition before finally being destroyed.

In the Inverell area of New South Wales, where the average annual rainfall is only 30″ and therefore not enough to promote deep weathering today, Tertiary relict soils have a marked effect on the nature of the modern soils on both basalt and granite, the dominating rocks in the area. Most plateau remnants are capped by lateritic soils, with a very hard capping of laterite remaining on some ridges. Other ridges have lost their lateritic cappings and the deeply weathered mottled horizons have been exposed to soil formation under the present environment. Both basalt and

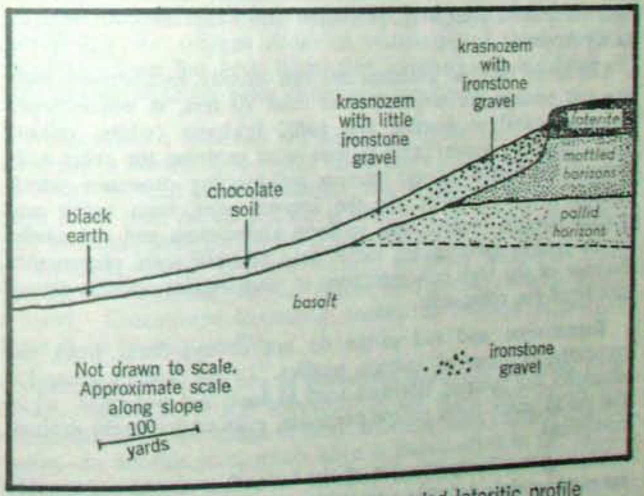

Figure 12.2: Soil formation on an eroded lateritic profile near Inverell, New South Wales.

granite have been laterised in this area, basalt tending to form more massive and vesicular laterite with poorly developed pallid horizons, and granite tending to form pisolitic laterite with deep, well-bleached pallid horizons containing secondary silica. Some ridges in the area contain krasnozems rather than lateritic soils, and it is difficult to determine whether these are the original soils, or whether they are the result of post-Tertiary modification of truncated lateritic profiles. Ridges which are below the general level of the plateau surface, and have presumably lost any Tertiary relict soils they may have contained, tend, during present weathering, to form shallow red soils without ironstone gravel (see Chapter 10). Because the Tertiary relict soils tend to be preserved on the higher ridges, soil depth tends to decrease down concave slopes—the opposite to what would occur if the soils were monogenetic. Fig. 12.2 illustrates a frequently-occurring catena in the Inverell area.

The lateritic profiles in the Inverell area have undergone considerable post-Tertiary modification, so the surface exposures are probably quite different from what they were like during the Tertiary. Because the sites are now freely drained, mottled and pallid horizons give rise to deep krasnozems lacking the bleached and yellow colours of the old profiles. The laterite itself has been strongly eroded, and is frequently found along the slopes as boulders and stones which are a nuisance to the wheat-growers who, to permit ploughing, pile them into heaps around trees or along fences.

Unaltered lateritic horizons are not usually encountered until the soil profiles are augered to at least 10 feet, at which depth colours typical of mottled and pallid horizons (white, yellow, blue) begin to appear; that is, one must examine the relict soils beyond the depth of the present soil-forming processes which modify the old profiles. In the Inverell area, both basalt and granite lateritic profiles tend to form krasnozems and red earths in the present environment, rather than podzolic soils, presumably because of the high concentrations of sesquioxides and the clayey nature of the relict soils.

Krasnozems and red earths do not always form from the truncated horizons of lateritic profiles. In the Sydney area, for example, the mottled horizons tend to form red podzolics, while the pallid zones form yellow podzolics, even under freely drained conditions.

During the Tertiary, most of the Inverell area was presumably mantled by deep lateritic soils, krasnozems and red earths. In the Late Tertiary to Pleistocene, the Kosciusko uplift raised the

Eastern Highlands, rejuvenating streams and accelerating erosion of these relict soils. A great deal of material eroded from these soils was carried west by the streams flowing towards the Darling and deposited as reddish, kaolinitic alluvium which underlies the black soil plains from Moree to Narrabri.

Following the Tertiary, soil forming on the freshly eroded basalt surfaces was black rather than red, so that the newer alluvium which was carried west was black and montmorillonitic, forming the black soil plains described in Chapter 10. Mineralogical evidence supports these origins for the two distinct alluvia of the north-western plains. The older, lower alluvium contains mainly resistant minerals indicative of rather severe chemical weathering, while the newer, upper alluvium contains many readily weathered minerals such as pyroxene and feldspar, indicating fairly weak chemical weathering of the basalt.

Eroded lateritic profiles are common along the northern and eastern Australian coasts and in the southwest of Western Australia. Many profiles are poorly preserved and it is extremely difficult to find profiles containing all three horizons (indurated, mottled and pallid) which have unequivocably formed by *in situ* alteration. Many of the so-called *in situ* profiles have formed on lateritic detritus, or the indurated horizon has been re-worked because it often contains extraneous material such as rounded gravels. Because of these factors, the pedologist has difficulty ascertaining the original depth of weathering of these profiles, but so far none has been found that compare with the depth of 200 feet recorded for some of the arid relict soils.

Relict lateritic soils and red earths are found on varied parent materials in Australia, but relict krasnozems are far more common on heavy textured parent materials such as basalt. Relict krasnozems have been recognised for a long time in eastern Australia (for example, Bryan, 1939) but their relationship to lateritic soils is not clear. Some krasnozems have probably weathered from lateritic soil horizons, whereas others were possibly always freely drained. Krasnozems frequently contain an horizon of laterite somewhere in their profile.

Silicification, though not as prominent as it is in the relict soils of arid Australia, is still important in the relict soils of humid areas, but silica illuviation is more common than surface silicification. In lateritic soils, much silica is re-deposited in the pallid horizons or just above them to form the hardened layer referred to in early soil literature as 'grey billy', although the term silcrete is preferred nowadays. Silcrete is not found in all lateritic profiles and is more common in those on silica-rich parent materials, such as granite or sandstone.

Both the nature of the lateritic horizons and the new environment of soil formation are important in the morphology of relict soils of humid areas. In humid areas it is rare to find Tertiary profiles which have been only slightly modified since their formation: they are more likely to serve as parent materials for soils under a later environment. In most of humid Australia, there is a tendency towards podzolisation, but if the exposed lateritic horizons are very high in iron, or very clayey (especially the ferricrete and mottled horizons), podzolic soils do not form, and iron- and clay-accumulating soils such as krasnozems and red earths form instead, even though such soils do not form today from fresh rock in the same area.

Tertiary Relict Soils in Arid Areas (Silcrete Profiles)

In Australia, the most obvious differences between the Tertiary relict soils of humid areas and those of arid areas are the composition of the duricrust, the depth of weathering, and the post-Tertiary modifications to the profiles. Arid relict soils tend to be capped by silcrete rather than ferricrete, though ferricrete is associated with some profiles (Mabbutt, 1965, found laterite grading into silcrete in some profiles in central Australia). Silcrete is just as variable in structure as laterite is. Some silcretes are massive, others contain cemented fragments of quartz and other minerals in a matrix of crystalline silica, and others are pisolitic. All silcretes are extremely hard; indeed, they are much harder than laterite, and hammers are sometimes broken when collecting samples from them. Since pure silica is colourless, many silcretes are whitish or greyish, but some are coloured red by very small amounts of iron oxides which may coat the quartz grains. This reddish colour has led some workers to confuse silcrete with laterite though the iron content of some reddish silcretes may be as low as 1%.

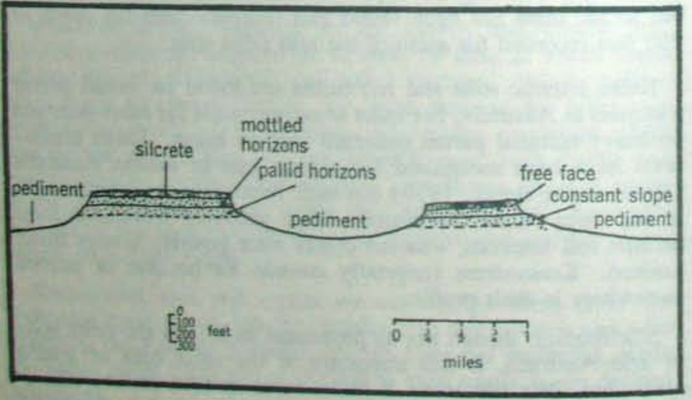

Figure 12:3: Duricrusted mesas in the Australian arid zone showing new pediment surface developed in bedrock below the deeply weathered profile.

The silcrete duricrust which caps the deeply weathered profile tends to protect the mottled and pallid horizons from erosion. The duricrust forms mesas and buttes from the Tertiary landsurface. The top of these landforms is very stony, but often quite flat. The duricrust forms the free face at the side of the mesas and buttes, while the less-resistant mottled and pallid horizons form the gentler constant slopes as shown in Fig. 12.3. The Tertiary deeply weathered landsurface is being eroded by backwearing, and the new pediment is frequently cut into bedrock, having completely stripped the relict soil profile. Some pediments are cut in the pallid horizons.

The pediment surface is usually strewn with silcrete boulders and stones eroded from the tops of the mesas and buttes. These boulders and stones decrease in size with increasing distance across the pediment from the duricrust capping. Soils formed on the pediment surface, being stone covered, are protected from deflation. The properties of these soils vary greatly beneath the stones, but often lime and gypsum accumulate somewhere in the profile. If the pediment is cut in the pallid horizon of the relict soil rather than in bedrock, the *stony desert soils* may inherit an acid reaction, for lime does not always invade the relict soil profiles, especially if the material upslope is also acid. If lime does invade the relict soil profile, the resulting soil is called a *calcareous lateritic soil.*

Other important soils associated with the Tertiary silcrete relict soils are the *red and brown hardpan soils.* The hardpan in these soils is siliceous and suggests that silification is still active in some parts of the desert. Since temperatures are very high in summer and rain occurs as torrential downpours rather than gentle drizzles, tropical conditions prevail for the few days in which rain falls. As there is a great deal of silica available on the pediments beneath silcreted residuals, it would be expected that under these conditions some of it could be illuviated to form a hardpan beneath the surface of these soils. Their profile is a fairly simple one, consisting of an undifferentiated horizon up to four feet deep, reddish or brownish in colour, resting on the silica hardpan which forms a perched water table enhancing deposition above it. The surface of the soil is usually overlain by silcrete gibbers. These soils are only partly polygenetic because, though the silica in the hardpan could be deposited during the present cycle of weathering and soil formation, it probably originated in the silcrete duricrust of the relict soils with which the red and brown hardpan soils are always associated.

The early idea (of Woolnough, 1927) that the deeply weathered Tertiary relict soils formed on a stable, flat landsurface close to base level is now unpopular. Recent evidence suggests that many of the deeply weathered relict profiles have developed on surfaces that were initially sloping rather than on surfaces that have been uplifted since the soils formed. However, some of the duricrusted profiles have undoubtedly been folded and faulted since their formation. A great deal of work is now in progress on the stratigraphic relations of duricrust and its chemical character, so the next few years should produce much useful information on Australia's Tertiary relict soils.

SOILS OF THE RIVERINE PLAIN

The Riverine Plain of southeastern Australia was defined by Butler (1958) as 'the region of plains in south-western New South Wales and north central Victoria which is traversed by the Murray, Murrumbidgee, Goulburn, Campaspe and Loddon rivers and their tributaries and distributaries.'

Much of the work in this area has been concentrated in the Murrumbidgee area, extending as far north as the Lachlan. The soils of the Riverine Plain have been the seat of controvery since 1950 when Butler first proposed that the plains were not built up by the present river system, but by an older system of rivers which no longer exist. These older rivers are called *prior streams*. The soils of the Riverine Plain form a complex succession on alluvial and loessial parent materials and have been studied by several pedologists. The main controversies are between Butler and Langford-Smith, and are concerned with the identification of the parent materials of the soils (whether they are wind-blown or riverine in origin), the number of depositional phases on the plain (whether there was one major alluvial and aeolian accession or several of each), and the nature of the depositional environments (whether they were arid or pluvial).

It was mainly from work in the Riverine Plain that Butler formulated his (1959) well-known *K*-cycle theory which was outlined in Chapter 5. Butler (1958) recognises three main depositional phases in the Riverine Plain, phases separated by stable periods of soil formation, but Langford-Smith's carbon-14 datings do not corroborate Butler's views. Since both Butler and Langford-Smith have published much detailed information on the soils of the Riverine Plain, it is not possible to do justice to all the literature here, so only broad generalities are presented. Students interested in the details of the controversies may consult the references given at the end of this chapter. The purpose of this section is to demonstrate soil stratigraphical procedures and to describe some of the soils of the area, rather than to interpret the soil data in terms of previous climates. To simplify the terminology, Great Soil Group names will be used, Butler's soil stratigraphical names being avoided where possible.

The Riverine Plain has been built up by accessions of material, both alluvial and aeolian, which pre-date the present stream system. The gradient over most of the plain is only slight, averaging one foot per mile, but it increases towards the Eastern Highlands, from which the streams debouch. The present streams trend southwest, whereas the prior streams trended west. Some of the courses of the prior streams can be identified from aerial photographs, for they are well preserved, showing little infilling since the streams abandoned them. The present river system uses some of the prior stream channels: on aerial photographs the two systems sometimes merge into each other. Some of the prior stream channels have been buried beneath later alluvium, but others can be traced by textural variation in the alluvium of the plain; the alluvium is sandy near the prior stream levees and becomes finer across the floodplains.

Soil distribution on the Riverine Plain bears little relationship to the present stream systems, but it is very closely related to the prior stream alluvium which controls soil texture, salinity, calcareousness and properties such as structure and *pH* which depend on these other properties. Since the levees of the prior streams are sandy and higher than their floodplains, red-brown earths tend to form on them. The lower, poorly drained floodplains away from the prior stream levees give rise to brown soils of heavy texture and grey soils of heavy texture, the latter forming in the most poorly drained areas, as shown in Fig. 12.4. With increasing distance from the prior stream, clay and salt content clearly increase, and sand and lime content decrease.

In the field it is difficult to distinguish between the alluvial and aeolian parent materials of the soils of the Riverine Plain, especially when looking at individual profiles. The only way to distinguish with certainty between the two is to study the pattern of variation in soil properties along transects. Soils on alluvial parent materials vary in properties along a line at right angles to the prior stream courses, as shown in Fig. 12.4, but soils on aeolian parent material vary in properties with distance from the source of the material.

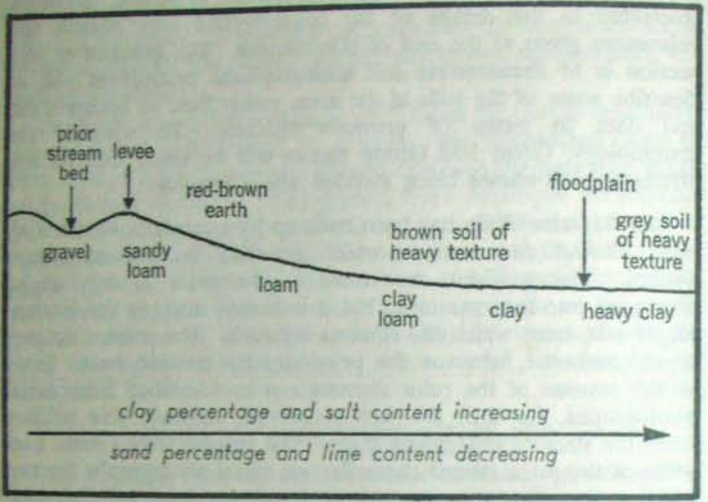

Figure 12.4: Section through soils formed on prior stream alluvium in the riverine plain, showing variation in soil properties from the stream across its floodplain.

Butler calls the aeolian material which blankets part of the Riverine Plain *parna* rather than loess, as it differs somewhat in properties from loess described from the Northern Hemisphere. He claims the parna originated to the west of the Riverine Plain in soils stripped of their vegetation during arid phases, when vast quantities of wind-borne dust moved across the plain from west to east in the path of the prevailing winds. Butler claims there were several accessions of parna, but the most ubiquitous one was the Widgelli.

Variation in the properties of parna occurs over a much greater distance than does variation in properties of the prior stream alluvium. The clay content of parna increases and the lime content decreases with increasing distance from the parna source (that is, with distance east). However, several tens of miles must elapse before the difference is noticeable; therefore, over short distances, soils on parna should reflect only site variations rather than parent material variations; in other words, parent material can be considered essentially uniform over short distances. Alluvial parent material, on the other hand, shows marked changes in texture and in salt and lime content over a few hundred yards.

Parna is frequently mixed with dune material consisting of saltation sand, adding further problems to the identification of its pattern of variation from its source. It is therefore not surprising that some pedologists have confused the alluvial and aeolian soils in the area. Since parna is fairly clayey away from the dunes, the red-brown earths which form in well-drained parna sites tend to be heavier-textured than the red-brown earths forming on the prior stream levees. Grey and brown soils of heavy texture occur on poorly drained parna sites, as well as on the prior stream floodplains.

The above comments refer to lateral variation in soil properties across the Riverine Plain. Vertical variation is even more striking, for the soils occur in stratigraphic succession according to the accessions of their parent materials. The type of relationship between the soils is shown in deep soil exposures.

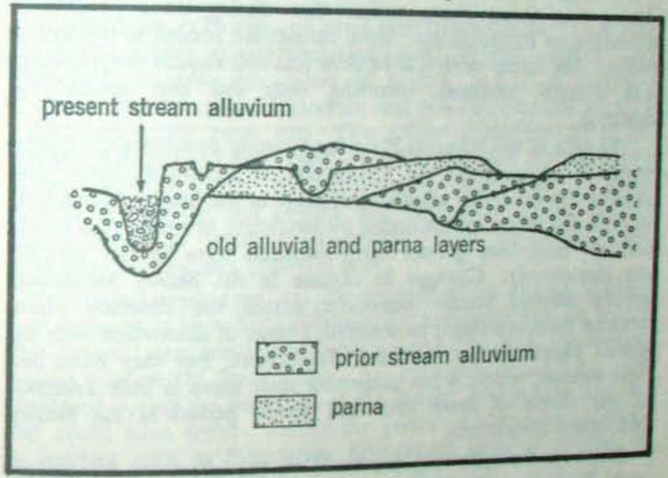

Figure 12.5: Diagram showing the relationship between alluvial and parna layers in the riverine plain.

The diagram shows that the prior streams have incised into the earlier materials, while the parna has blanketed the landscape, without incision. The buried soils have in many cases been modified, either by truncation or by the processes which developed the newer soils on the materials which overlie them. The section shown in Fig. 12.5 clearly illustrates Butler's (1959) concept of *groundsurfaces* as being the alternation between periods of erosion and deposition (when the soil parent materials were deposited) and periods of soil formation.

The material presented in Fig. 12.5 involves a certain amount of interpretation of field soil data; for example, some workers may question the aeolian origin of the parna and claim that this material, also, may be riverine in origin. However, relating this section to past climates demands even larger assumptions. Butler favours deposition of the prior stream alluvium during arid phases, when stream discharge was less and the streams were unable to carry their load of sediment. Langford-Smith, on the other hand, favours deposition during pluvial phases, and he claims, from measurements of meander belts of the prior streams, that these were considerably greater than the present rivers—having, indeed, discharges comparable with those of the present Mississippi. Langford-Smith's views are supported by carbon-14 datings which suggest that most of the prior stream alluvium was deposited during one major phase, rather than during three periods of accession, as Butler claims. More datings are needed to clarify the picture, but future work will be slow and will require deep borings and detailed fieldwork involving both soil and geomorphic evidence.

The soils of the Riverine Plain have mostly developed on parent materials deposited in the Pleistocene or Recent Periods. The streams depositing the alluvium originate in the Snowy Mountains, so their discharges are increased by meltwater in spring and would certainly have been greater when snowfalls were heavier (during the glaciations). Changes in climate in the Snowy Mountains greatly affected stream behaviour across the Riverine Plain; workers have attempted to correlate phases of alluviation with the glacial phases of the Northern Hemisphere, but they have had little success, which is not surprising since there is little evidence of the effects of more than one glacial period in the Snowy Mountains themselves.

Much of the following discussion on the effects of the Recent Arid Period on soils is relevant to the soils of the Riverine Plain, for some of the upper soils in the stratigraphic sequence are post-Pleistocene. Some of the aeolian parent material was probably deposited during the Recent Arid Period when vegetation was depleted and the soils to the west of the Riverine Plain were exposed to wind-erosion.

SOILS OF THE RECENT ARID PERIOD

The effects of previous aridity are less well preserved in soils than are the effects of previous greater rainfall, because the materials accumulating during the aridity are later leached. However, some soils do have properties indicating polygenesis, with the earlier environment being the one when there was less leaching. The solodic soils discussed in Chapter 11 are examples. Degraded chernozems, common in the United States, but not very common in Australia, provide another example.

The profile of a degraded chernozem is essentially that of an ordinary chernozem, with the effects of weak podzolisation showing in the upper horizons; often, a weak *E* horizon (bleicherde) has developed, the upper horizon is more acid than the lower horizons, and there may be some illuviation of clay. Many of the chernozems of the eastern part of the chernozem zone in the United States are undergoing degradation under the present environment, suggesting that their chernozem characteristics were the products of a climate drier than the present one, which is less conducive to chernozem formation, and favours podzolisation instead.

In Australia, both soil and plant distributions point to a period of greater aridity in the immediate past. Several workers have attempted to date this increased aridity, and it is generally accepted that it is post-Pleistocene, and possibly occurred between 4,000 and 6,000 years ago. This aridity is referred to as the *Recent Arid Period* by Downes (1954). It has had two main effects on soils. First, the devegetation induced by the lower rainfall accelerated soil erosion and deposition, so old profiles were buried, in many cases beneath fresh parent materials, especially aeolian and colluvial ones.

In sloping areas (e.g., on the western slopes around Mudgee and Bathurst in New South Wales), the last groundsurface before the present one frequently contains colluvial soils resembling prairie soils on the lower slope. The colluvium could have been emplaced during the beginning of the Recent Arid Period, and the soil could have formed during the period of decreased rainfall, as modern soils in these areas have much stronger features of podzolisation.

Near the Nepean River in the western part of the Cumberland Plain surrounding Sydney there is an area of fixed longitudinal sand dunes, resting on alluvium which is thought to be Late Tertiary in age (Simonett, 1950). The soils on these dunes are very different from surrounding alluvial soils and support different vegetation and land-use. Simonett considers that these dunes were formed from eroded soils on the sandstone plateau of the Blue Mountains to the west, during the Recent Arid Period when the plateau was denuded of vegetation. Aeolian deposits such as these dunes are common in areas to the east of sources of loose material.

Pedogenic Zone	Average Annual Rainfall		Salt Accumulation		Post-arid Leaching	Dominant Soil-Forming Processes
	Arid	Post-arid	Arid	Post-arid		
1	>22	>40	none	none	very high	podzolisation
2	16-22	30-40	slight	none	high	slight solonisation then podzolisation
3	11-16	20-30	considerable	none	moderate	solodisation
4	8-11	15-20	slight	slight	slight	solonisation or slight solodisation
5	<8	<15	very slight	considerable	little or none	solonisation

Figure 12.6: Downes' (1954) five pedogenic zones indicating the effects of cyclic salt on polygenetic soils in south-eastern Australia.

Lunettes, which are crescentic deposits of fine material common on the lee-side of lake beds in semi-arid parts of southern Australia, are thought to have formed in a similar manner.

The second important effect of the Recent Arid Period on soils is the salt accession which affected all soils forming during this period. Thus any soil which is known to be older than 6,000 years would have contained more salts than it does now. Downes (1954) postulated that during the Recent Arid Period rainfall in south-eastern Australia was about half the present amount, enabling cyclic salt (see Chapter 11) to accumulate in soils from which present-day leaching largely removes salts. That is, the soils underwent salinisation during the Recent Arid Period, but have been subsequently desalinised as described in Chapter 11. This theory explains the anomalous distribution of soils within the podzolic zone, where the most highly differentiated profiles are often found in the driest zones. Downes divided the soils of south-eastern Australia into five pedogenic zones related to the present rainfall (based on the assumption that rainfall in the Recent Arid Period was about half the present amount). His pedogenic zones are shown in Fig. 12.6.

The point was made in Chapter 11 that salt accumulation in a soil depends not only on leaching, but also on the rate of accession from the atmosphere, that is, on the rate of accession of cyclic salt. In areas of very low rainfall, accession is very slight, and, since rainfall in arid areas comes in heavy falls rather than gentle drizzles, the cyclic salt is leached as it falls. In very humid areas, leaching exceeds the rate of accumulation of cyclic salt. For these reasons, there is only slight or insignificant cyclic salt accumulation in zones 1, 2, 3 and 4 under the present rainfall (under the heading 'post-arid' in Fig. 12.5), and in zones 1, 2, 4 and 5 under the rainfall of the Recent Arid Period (under the heading 'arid' in Fig. 12.5).

Downes found the greatest cyclic salt accumulation in soils close to the 12″ average annual isohyet under the present climate and the 14″ average annual isohyet during the Recent Arid Period. Thus maximum solonisation occurred in zone 3 during the Recent Arid Period and occurs in zone 5 today. Solodisation, which is solonisation followed by podzolisation, is therefore maximum in zone 3. In zone 2 podzolisation operated weakly even in the Recent Arid Period when there was only slight salt accumulation, but in zone 4 podzolisation in the present climate is still weak, so solodisation is slight. One would therefore expect solodic soils to be most common in areas receiving between 20″ and 30″ rainfall per annum, and this is certainly so.

The accession of cyclic salt affected all soils in southeastern Australia, in areas receiving between 15″ and 40″ rainfall per annum, excepting those with a high calcium-status, that is chocolate soils, black earths, terra rossas and rendzinas. For this reason many pedologists regard most soils in this area as polygenetic (provided the landsurface has been stable for a considerable time) though there may be little field evidence for their polygenesis. Thus many of the soils shown on Australian soil maps to be podzolics are really solodics.

The study of relict soils in Australia is providing a great deal of information on previous climates and landforms; however, there is much more work to do as the field of study is a relatively new one and it is quite likely that the work of the next few years may necessitate changes in many of the comments in this chapter.

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Appendix

Some Laboratory Estimations of Soil Properties

1. A Quick Method for the Mechanical Analysis of Soils

There are two main steps in the mechanical analysis of soils. First, the soil has to be separated into individual particles, because, in the field, it occurs as aggregates that are cemented by clay. This separation into individual particles is called *dispersion* and is achieved by adding a dispersing agent (a sodium compound) and water to the soil and mixing them. The second step is the *fractionation* of the individual soil particles into gravel, coarse sand, fine sand, silt and clay. The coarser particles (gravel and sand) are separated by sieving, but the silt and clay fractions are usually separated in water. The method of mechanical analysis, selected here for its speed, uses motor dispersion and hydrometer fractionation; that is, the soil is dispersed by agitation with an electric mixer, and the silt and clay are estimated by recording with a hydrometer the density of the soil-water suspension.

The method presented here is based on Bouyoucos' (1936) method. Although this method may not be as accurate as some others, it is very quick (more than a dozen samples can be analysed in a day) and accurate enough considering that samples of soils collected in the field only a foot apart may give quite different results. The equipment required is available in most laboratories. There are two exceptions, the soil hydrometer which is calibrated to read directly in grams of soil colloid per litre at 67°F, and the electric mixer. Milk-shake mixers, high-torque stirrers or kitchen mixmasters are all suitable for motor dispersion.

If the soils to be analysed contain appreciable quantities of organic matter or lime (more than 1% of each), these constituents should first be removed. Lime[1] may be removed by adding dilute (4N) hydrochloric acid to the soil until effervescence ceases.

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[1] The hydrochloric acid treatment really removes carbonate minerals from the soil. The most common form of carbonates in soils is lime. This method gives a rough estimate of the percentage carbonate in the soil. Weigh the sample before and after acid treatment, the difference being the weight of carbonate in the soil. Convert this to a percentage relative to the total soil (i.e., the first weight).

Organic matter may be removed by boiling the soil in hydrogen peroxide. Most Australian soils, however, can be mechanically analysed without these treatments.

The results are usually given as the percentage sand, silt and clay in the non-gravel fraction of oven-dry soil, and gravel is given separately as a percentage of the total soil. In the following directions, International (Atterberg) size gradings are used (see Chapter 2).

Dry about 110 gms of soil in an oven at 105°C for at least eight hours. Sieve 100 gms of the oven-dry soil through a 2mm (10 mesh) sieve. Weigh[2] the gravel fraction remaining in the sieve and record it separately as a percentage. Place 60 gms of sieved soil into the beaker for the electric mixer. Add 50 mls of 5% calgon[3] brought to *pH* 8.5 with sodium carbonate, using a *pH* meter. Fill the beaker to 2/3 with distilled water and agitate the contents with the mixer for 10 minutes. Wash the contents of the beaker into a litre measuring cylinder and fill the cylinder to the 1,000 ml mark with distilled water. Place a rubber stopper over the end of the cylinder and shake the contents vigorously for one minute. Allow to settle, noting the time when shaking ceased.

After five minutes, measure with the soil hydrometer the grams of soil colloid in suspension, and record the temperature of the suspension. Remove the hydrometer from the suspension tc prevent soil particles from settling onto it.

After two hours, repeat the hydrometer and temperature readings. The five-minute hydrometer reading gives the amount of both silt and clay in suspension, whereas the two-hour reading gives the amount of clay in suspension, as the silt settles to the bottom during the two hours. It takes approximately five minutes for the coarse and fine sand to settle through a distance of 10 centimetres.

Pour off the supernatant clay suspension from the cylinder and wash the settled sand and silt into a beaker. Fill the beaker to about 10 centimetres, stir and allow the sand to settle for five minutes. Pour off the supernatant silt suspension and repeat the procedure until the supernatant liquid is clear (three or four decantations are usually enough). Dry the washed sand in an oven at 105°C.

Sieve the dry sand through an 0.2 mm (100 mesh) sieve. Weigh each fraction (the fraction remaining in the sieve is the coarse sand and the fraction passing through is the fine sand).

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[2] All weights, and hydrometer and temperature readings need be taken only to one decimal place.

[3] Calgon, sodium hexametaphosphate, is available from grocery stores.

Correct the hydrometer readings as follows:

1. for temperature—add 0.3 gm for every 1°C above 19.5C° and subtract 0.3 gm for every 1°C below 19.5°C.
2. for weight of calgon and sodium carbonate—subtract 0.3 gm.

The results may be set out in tabular form as follows:

PERCENTAGE GRAVEL IN SOIL:..

Time since shaking	Hydrometer reading in gms soil colloid/litre	Temperature °C	Hydrometer correction in gms	Corrected hydrometer reading in gms
5 minutes (silt plus clay)				
2 hours (clay)				

	Non-gravel fractions	
	Gms	Percentage
clay		
silt		
fine sand		
coarse sand		
Total		

The results rarely add to exactly 100% or 60 gms (the weight used at the start). It is more common to obtain more than 100% as the fine sand and silt sometimes overlap in the fractionation, that is, some fine sand is estimated in the five-minute hydrometer reading, and some silt is weighed with the fine sand.

2. Some Experiments on Soil Water

Soil moisture content is conventionally expressed in terms of oven-dry soil. Various points along the soil's moisture curve (Fig. 3.3) may be estimated roughly as follows:

Field capacity: saturate a soil in the field and cover it with plastic or hessian for 48 hours. Collect a sample in an air-tight container. Place about 30 gms into a small beaker and weigh the beaker and soil. Oven-dry the soil at 105°C for 48 hours and weigh the soil and beaker again. The loss in weight represents the moisture content at field capacity. Convert it to a percentage of the oven-dry soil as shown in Chapter 3.

Permanent wilting point: grow a plant (e.g., a lettuce) in a pot containing the soil. Water the soil regularly until the plant is fairly robust, and then allow the soil to dry until the plant wilts. Re-water the soil, and the plant will revive. Repeat the process until the plant fails to revive. A sample of the soil should be collected in an air-tight container each time before the soil is re-watered: if the plant revives, replace the sample in the pot. Plants usually undergo three or four temporary wiltings before the permanent wilting point is reached. Oven-dry the sample as for field capacity determinations, and calculate the moisture content of the soil as a percentage of the oven-dry soil at permanent wilting point.

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GLOSSARY

The definitions given for the terms which follow are those used in this book, and not necessarily those of various authorities. Where possible, traditional usage is given, but Australian usage is given where traditional usage is no longer popular.

The definitions are as simple as possible and students are referred to the text, via the index, for more complete explanations. Not every term used is defined in the glossary; for example, terms used in the Seventh Approximation (Guy Smith, 1960) and FAO/UNESCO Soil Classification (1968) are not re-defined in the glossary since they are defined in Chapter 6 and not used elsewhere in the book. Not every Great Soil Group or soil-forming process is re-defined: the index should be used instead. Traditional Great Soil Groups which are constantly referred to in the early chapters before they are fully discussed in their context are, however, defined in the glossary, e.g., podzol, chernozem.

Rock definitions are kept to a minimum: students should refer to any elementary geology text for rock names. Mineral definitions are greatly simplified and students should also use a geology book for more information. Self-explanatory terms such as tetrahedral, and non-technical terms such as cement are not included in the glossary.

Chemical elements are not included in the glossary but the common names of some chemical elements are included, e.g., halite, gypsum, borax and alumina. Students should consult an elementary chemistry text for the properties of uncommon elements such as titanium and molybdenum.

ABSORB: To take into, like salt into water.

ADSORB: To be attracted to or to adhere to, like cations to a clay micelle or nucleus, or iron filings to a magnet.

AEOLIAN MATERIAL: Any wind-transported material: includes loess which is moved in suspension, and dune sands which move by saltation.

AEROBIC: Freely drained, fully aerated, oxidising conditions.

AIR CAPACITY OF A SOIL: The quantity of air, expressed as a percentage of the total soil constituents, remaining in a soil after it has been draining for 48 hours, following thorough wetting (in Australia).

ALBITE: Soda-plagioclase (see Plagioclase).

ALKALI CATIONS: Group 1 of the periodic table of chemical elements; in soils the most common are sodium and potassium.

ALKALINE EARTHS: Group 2 of the periodic table of chemical elements: in soils calcium and magnesium are the most common.

ALLUVIUM: Material transported and deposited by streams.

ALUMINA: Aluminium oxide.

ALUMINOSILICATES: The main rock-forming minerals of the earth's crust: these weather to form the main minerals in soils.

AMINO ACIDS: Organic acids containing NH_2.

AMMONIFICATION: The conversion of plant and animal remains into ammonia by soil micro-organisms.

AMORPHOUS: Non-crystalline: no definite pattern of arrangement.

AMPHIBOLE: A group of readily weathered aluminosilicate minerals with double chain structures: the chains are linked by magnesium, iron, calcium and sodium.

ANAEROBIC: Poorly drained, poorly aerated, reducing conditions.

ANATASE: Titanium oxide.

ANDESINE: Sodic-calcic plagioclase (see Plagioclase).

ANHYDRITE: Calcium sulphate ($CaSO_4$).

ANION EXCHANGE: The replacement of anions on clay micelles by other anions.

ANIONS: Negatively charged elements formed by the addition of electrons, leaving a deficit of positive charge on the nucleus.

ANORTHITE: Calcic plagioclase (see Plagioclase).

ANTECEDENT STREAM: A stream, the course of which was established prior to orogeny or uplift.

APATITE: A calcium phosphate mineral containing some fluorine, and chlorine or hydroxyl: may be a primary mineral or form in the soil from phosphates.

APEDAL: The type of soil structure lacking peds, or natural structural units roughly equal in size and shape: there are two types of apedal soils—single grain or massive.

ASPECT: The direction a slope faces.

ATTERBERG SCALE: The internationally accepted system of size gradings which is a logarithmic scale as follows: gravel more than 2.0 mm; coarse sand 2.0-0.2 mm; fine sand 0.2-0.02 mm; silt 0.02-0.002 mm; and clay less than 0.002 mm diameter.

AUGITE: One of the most common pyroxenes (see Pyroxene).

AUTHIGENESIS: The formation of new minerals during the consolidation of sedimentary rocks or subaerially in the regolith.

AZONAL SOILS: Soils which lack profile differentiation, because of immaturity or because erosion is faster than soil formation, preventing horizons from forming.

BASALT, DOLERITE: Fine-grained, basic igneous rock consisting mainly of pyroxene, plagioclase and olivine.

BASE EXCHANGE: The replacement of basic cations on the clay micelle by other basic cations: the most common soil bases are calcium, sodium, potassium and magnesium.

BASES: The cations of the alkali metals and alkaline earths; the most common bases in soils are calcium, sodium, potassium and magnesium.

BEIDELLITE: A member of the expanding lattice group of clay minerals with two silica and one alumina sheet forming the crystal lattice.

BIOTITE: Black mica (see Mica).

BLEICHERDE: The bleached horizon formed beneath the soil surface by the eluviation of constituents to the subsoil: the E horizon.

BOEHMITE: A hydrated oxide of aluminium found in soils.

BORAX: Sodium borate: the common form in which boron is added to the soil to make up a deficiency.

BYTOWNITE: Calcic-sodic plagioclase (see Plagioclase).

CALCITE: Crystalline calcium carbonate: may be a primary mineral, or form in the soil from carbonates.

CALCRETE: Any material hardened by enrichment by calcium carbonate.

CAPILLARITY: The process whereby water moves by surface tension through extremely small pores.

CAPILLARY CAPACITY: See field capacity.

CAPILLARY FRINGE: The height above the water to which the soil's capillary pores remain saturated.

CATENA: The succession of soils along a slope.

CATION EXCHANGE: The replacement of cations on a clay micelle by other cations.

CATION EXCHANGE CAPACITY: The ability of a clay to exchange cations from the soil solution: abbreviated to C.E.C.

CATIONS: Positively charged elements formed by the removal of electrons, leaving excess positive charge on the nucleus.

CHELATES: Complexing organic agents in the soil having the ability to combine with metallic cations.

CHERNOZEM: A soil found in sub-humid grasslands: it is usually high in silicate clay minerals with minimal clay illuviation and a dark-coloured A horizon containing at least 4% organic matter: lime is illuviated to the B horizon.

CHLORITE: A member of the hydrous mica group of clay minerals with two silica and one alumina sheet forming the lattice.

CLAY: Mineral particles defined on the International (Atterberg) scale as having a diameter less than 0.002 mm and which remain in suspension by Brownian motion when shaken up in water.

CLAY SKINS: Shiny surfaces on wet soils caused by the deposition of clay, often by illuviation.

CLIMATIC CLIMAX VEGETATION: Vegetation associated with a mature soil in dynamic equilibrium with its environment.

CLODS: Soil aggregates caused by cultivation, especially ploughing.

COLLOIDS: Particles which remain in suspension by Brownian motion when shaken up in water: they are defined internationally as particles smaller than 0.002 mm diameter: may be humic and mineral.

COLLUVIUM: Material which has moved downslope by gravity and running water: any of the products of mass movement along slopes.

COMMON SALT: Sodium chloride.

CONCRETION: A rounded aggregate of segregated soil material that is concentric about a nucleus: a nodule that is concentric about a nucleus.

CYCLIC SALT: Salt added to the soil by rainwater.

DEUTERIC ALTERATION: Alteration of primary rock minerals by magmatic solutions at high temperatures during the later stages in the consolidation of rocks.

DIASPORE: See boehmite: boehmite and diaspore are optical isomers.

DICKITE: A mineral of the kaolin group (see Kaolin).

DIFFUSION: The process whereby gases of differing composition mix by spontaneous agitation until they form a single gas uniform in composition throughout.

DIORITE: A coarse-grained, intermediate, igneous rock consisting mainly of feldspar, amphibole and biotite.

DIPOLE: A two-poled substance such as water which has positive (the hydrogens) and negative (oxygen) ends.

DOLERITE: See Basalt.

DOLOMITE: Calcium-magnesium carbonate.

DOLOSTONE: A sedimentary rock composed mainly of dolomite (calcium-magnesium carbonate).

DUNITE: An ultra-basic igneous rock containing mainly olivine.

DUPLEX PROFILE: (Used in Northcote's system of soil classification.) A soil in which there is a marked increase in clay in the B horizon compared with the A horizon.

DURICRUST: An indurated surface horizon overlying a deeply weathered soil profile.

DYNAMIC EQUILIBRIUM IN SOILS: The state in which the processes of soil formation are just able to keep pace with accessions of soil material to maintain the soil profie in a constant form with the passing of time.

ELUVIAL HORIZONS: Horizons in the soil that have lost constituents by leaching to lower horizons in the profile.

ELUVIATION: The process of removal of constituents from the upper horizons of the soil profie: some of the eluviated constituents are deposited in lower horizons.

FALLOWING: The practice of leaving the soil free of crops and weeds for a period of time (usually not longer than 6 months) in order to conserve soil moisture by reducing transpiration losses.

FELDSPAR: A group of aluminosilicate minerals with a continuous framework of silicon tetrahedra: the tetrahedra are linked by potassium, sodium and calcium: the main feldspars are orthoclase (potash feldspar) and the plagioclases (sodic-calcic feldspars).

FELDSPATHOIDS: Aluminosilicate minerals similar in structure to the feldspars, but with a lower silica to alumina ratio: they form in silica-deficient magmas such as basalts and are very susceptible to subaerial weathering.

FERRALISATION: Soil-forming processes involving deep weathering under free drainage, and the accumulation of sesquioxides throughout the profile which has no distinct horizons as in laterisation.

FERRIC ION: Iron III, a cation formed by the removal of three electrons from the element iron.

FERRICRETE: Any material hardened by enrichment by iron.

FERROUS ION: Iron II, a cation formed by the removal of two electrons from the element iron.

FIELD CAPACITY: The amount of water remaining in a soil after it has drained for 48 hours (in Australia) following thorough wetting and covering the soil to prevent evaporation; approximates *pF* 2.5 or ⅓ of an atmosphere pressure.

FIXATION: The rendering insoluble or making unavailable, by elements or organisms within the soil, of a plant nutrient: phosphorus is fixated by iron and aluminium oxides; nitrogen is fixated by bacteria in the nodules on legumes for use by the host plant (see Symbiotic fixation).

FLOCCULATION: The moving together of soil particles leaving spaces between the floccules.

FOSSIL SOIL: See Relict soil.

FRAGIPAN: An horizon in a soil consisting of an irregular, very hard accumulation of acid clay thought to have formed by illuviation of clay down cracks in the profile: found in periglacial areas in material which cracks readily.

FULVIC ACID: That fraction of humus which is highly mobile but colourless.

GENETIC THEORY OF SOIL FORMATION: The theory which states that the soil horizons are related to each other and differ from the parent material because of the action of the environment.

GIBBSITE: The most common hydrated aluminium oxide found in soils: consists of a double layer of hydroxyl molecules enclosing aluminium in the octahedral position.

GILGAI: Surface undulations in the soil forming small ridges and hollows.

GLEYING: The process of weathering under reducing conditions due to horizon saturation: bleaching occurs through the migration of coloured ions from the gleyed horizon; any coloured ions remaining, being in the reduced form, tend to form blue, grey and green colours.

GLEY HORIZON: An horizon which has been water-logged for some time and has weathered under reducing conditions which promote bleaching and the development of blue, grey and green colours.

GNEISS: A foliated, high-grade, regionally metamorphosed rock frequently containing much feldspar and amphibole.

GOETHITE: One of the less hydrated iron oxides producing a reddish-brown colour in soils.

GRADITIONAL PROFILE (used in Northcote's system of soil classification): a soil in which there is a slight increase in clay content in the *B* horizon compared with the *A* horizon.

GRANITE: A coarse-grained, acid, igneous rock containing mainly quartz and feldspar and a little biotite and hornblende.

GRAVEL: Mineral particles defined on the International (Atterberg) scale as having a diameter greater than 2.0 mm.

GREAT SOIL GROUP: The level in soil classification based on total profile features: the general names for soil, e.g., podzol, red podzolic, chernozem and red-brown earth.

GROUND SURFACE: The term Butler uses for the soil and parent material formed during one *K*-cycle: see *K*-cycle.

GYPSUM: Hydrated calcium sulphate ($CaSO_4.2H_20$).

HAEMATITE: One of the least hydrated of the iron oxides, producing a reddish colour in soils.

HALITE: Sodium chloride or common salt.

HALLOYSITE: A hydrated clay mineral with one silica and one aluminum sheet forming the crystal lattice.

HORIZON: See soil horizon.

HORNBLENDE: One of the most common amphiboles (see Amphibole).

HUMIC ACID: That fraction of humus which is dark coloured and fairly immobile.

HUMIFICATION: The processes whereby fresh organic matter in soil decomposes into humus.

HUMIN: An insoluble fraction of humus.

HUMUS: That part of soil organic matter which has decomposed into a dark-coloured, soapy, amorphous mass without trace of the original plant tissue.

HYDRATION: The addition of water to an atom, ion, molecule or mineral.

HYDRONIUM: An ion formed by the addition of hydrogen to a water molecule (H_3O^+).

HYGROPHYTIC: Water-loving: hygrophytic plants require large quantities of water and have little resistance to drought.

HYDROPONICS: The growing of plants in nutrient solutions rather than in soil.

HYGROSCOPIC: The ability to take in water from the atmosphere.

HYDROXYAPATITE: See apatite: the type of apatite in which hydroxyl occurs instead of chlorine: it forms in the soil from superphosphate if lime is present. It is very insoluble.

HYDROXYL: An anion composed of oxygen and hydrogen (OH).

HYSTERESIS: The difference in moisture obtained when drying out and when wetting up a soil for the same pressure values.

IDDINGSITE: A group of clay minerals (smectite-chlorites) formed by the alteration of olivine either deuterically or subaerially.

IGNEOUS ROCKS: Rocks which have consolidated from molten magma.

ILLITE: A member of the hydrous mica group of clay minerals with two silica sheets and one alumina sheet forming the crystal lattice.

ILLUVIAL HORIZONS: Horizons in the soil that have received constituents leached from upper horizons in the soil profile.

ILLUVIATION: The process of deposition of constituents leached from upper horizons into lower horizons of the soil profile (cf. Eluviation).

INDURATE: To harden.

INTRAZONAL SOILS: Soils which do not reflect their climatic zone because of the dominating effect of some other factor of soil formation, especially relief and parent material.

IONS: Charged elements formed by the addition or removal of electrons: *cations* carry positive charges and *anions* negative charges.

K-CYCLES: Butler's scheme of naming soil successions back from the present in a periodic soil stratigraphical sequence.

KANDITES, KAOLIN: A group of clay minerals with single silica and alumina sheets forming the crystal lattice: the main members are kaolinite, nacrite and dickite.

KAOLINITE: A member of the kaolin group of clay minerals.

KRASNOZEM: A deep, red soil formed under free drainage in the humid tropics: it is clayey in texture and dominated by sesquioxides.

LABRADORITE: Calcic-sodic plagioclase (see Plagioclase).

LACUSTRINE: Pertaining to a lake: lacustrine deposition—deposition in the bed of a lake.

LATERISATION: Soil-forming processes occurring in the humid tropics and involving deep weathering, removal of bases and some silica, and the accumulation of sesquioxides in marked horizons in relation to a fluctuating water table.

LATERITE: Indurated, iron-rich material which hardens on exposure and is associated with a deeply weathered profile.

LATERITIC SOIL: A sesquioxide-dominated, deeply weathered soil containing a sequence of horizons of laterite over mottled over pallid zone, formed under a fluctuating water table.

LATOSOL: A term common in America for all red tropical soils, generally deeply weathered.

LAVA: Molten igneous rock extruded at the earth's surface: solidifies into a fine-grained, volcanic rock such as basalt.

LEACHING: The removal of material from a soil, by water percolating through the soil: the most commonly leached constituents are bases and plant nutrients: leaching involves the removal of constituents in solution, as ions, and in suspension, as colloids.

LEGUME: A plant which has the natural ability to fixate gaseous nitrogen directly from the soil air through bacteria in nodules on its roots, e.g., beans, clover, medicks.

LEUCOXENE: A hydrated titanium oxide.

LIME: Salts containing calcium: the most common form is calcium carbonate: calcium oxide and calcium hydroxide are also lime.

LIME REQUIREMENT: The amount of lime needed by a soil to bring its *pH* to about 6.5.

LIMNITE: One of the most hydrated iron oxides, producing a yellowish colour in soils.

LIMONITE: One of the more hydrated iron oxides, producing a yellowish-brown colour in soils.

LITHOSOL: A shallow, undifferentiated soil formed on rock.

LOAM: The soil texture class which is friable and coherent: the soil can be rolled into a ribbon, but not moulded into a ring: loams contain more than 50% sand, less than 25% silt and less than 35% clay.

LOESS: Material transported in suspension and deposited by wind: the particles are fine-grained, usually silt and clay sized.

MAGMA: Molten rock.

MAGNETITE: Black, unhydrated, magnetic iron oxide.

MATURE SOIL: A soil in dynamic equilibrium with its environment: see Dynamic Equilibrium.

MECHANICAL ANALYSIS: The laboratory assessment of a soil's particle size grading, by dispersing the soil into individual particles and dividing these into size fractions on the basis of their settling velocities, and by sieving.

MEDICK: A creeping legume widely used as a pasture (see Legume).

METAMORPHIC ROCK: Rocks which have crystallised from earlier rocks by deformation by heat and pressure.

MICA: A group of aluminosilicate minerals with sheet structure: the sheets are linked by potassium, sodium, iron and magnesium: main types are muscovite (white) and biotite (black).

MICELLE: The structural unit or nucleus of a clay mineral which behaves like a huge anion attracting cations and the positive ends of water molecules.

MICRO-RELIEF: Surface unduations in the soil (see also Gilgai).

MOISTURE CHARACTERISTIC OF A SOIL: A graph relating the amount of water in a soil to the pressure with which that water is held in the soil: the pressure is usually plotted as *pF* and the amount of water as a percentage of oven-dry soil.

MONTMORILLONITE: A member of the expanding lattice group of clay minerals with two silica and one alumina sheet forming the crystal lattice.
MULCH: A covering over the surface of the soil to lower water loss by free evaporation from the soil surface: may be natural (see Self-mulching soil) or artificial.
MURIATE OF POTASH: Potassium chloride: KCl.
MUSCOVITE: White mica (see Mica).

NACRITE: A mineral of the kaolin group (see Kaolin).
NITRIFICATION: The conversion of ammonium ions in the soil into nitrate ions.
NODULE: A rounded aggregate consisting of segregated soil material such as iron oxides or lime: nodules may be amorphous (no definite pattern of arrangement and non-crystalline) or concretionary: a rounded lump on the roots of legumes containing nitrogen-fixing bacteria.
NON-SYMBIOTIC FIXATION OF NITROGEN: The taking in of atmospheric nitrogen by organisms in the soil: these organisms exist independently of plants.
NUTRIENT: An element essential for growth.

OLIGOCLASE: Sodic-calcic plagioclase (see Plagioclase).
OLIVINE: A group of readily weathered aluminosilicate minerals in which the silicon tetrahedra are linked by iron and magnesium.
OPEN SYSTEM: A system in which materials can enter and leave without effecting any permanent change in the properties of the system.
ORGANIC MATTER: All the constituents of soil arising from living matter: includes plant remains and micro-fauna: may be fresh or decomposed (humified).
OROGENY: Mountain-building uplift.
ORTERDE: An accumulation of humus and sesquioxides in podzolised soils: thicker and less indurated than ortstein.
ORTHOCLASE: Potash feldspar (see Feldspar).
ORTHOSILICATES: Independent groups of silicon and oxygen tetrahedra linked by cations, e.g., olivine group, zircon.
ORTSTEIN: A thin, indurated horizon found in podzolised soils from the accumulation of sesquioxides and humus.
OVEN DRY: The weight of the soil after it has been in an oven at 105° C. for at least 48 hours: many laboratory determinations are based on oven-dry soil.
OXIDATION: The removal of electrons from an atom or molecule, e.g., $Fe^{2+} - e \rightarrow Fe^{3+}$.

PALAEOSOL: See Relict soil.
PALYGORSKITE: A silicate clay mineral belonging to the open, fibrous class.
PARALLELIPIPEDAL: The unit of soil structure consisting of skewed parallelograms.
PARNA: Wind-borne material found in the Riverine Plain of south-eastern Australia.
PEAT: An accumulation in the regolith in which the organic component is extremely high (usually at least 40%) in relation to the mineral component: soil formation is impeded by the excess organic matter in peats.
PEDALITY: The aspect of soil structure concerned with the size, shape, proportion and firmness of peds.
PEDOLOGY: The study of the genesis and morphology of soils, with particular emphasis on field soil profiles.
PEDS: The natural units of soil structure formed by the soil's tendency to fracture along regularly spaced planes of weakness.
PERIGLACIAL: Peripheral to areas that are permanently snow-covered.

PERIODICITY: The alternation of phases of soil formation with phases of erosion-deposition.
PERMANENT WILTING POINT: The amount of water remaining in the soil when plants fail to revive after wilting: approximately *pF* 4.2 or 15 atmospheres pressure.
pF: A convenient scale for measuring the pressure at which water is held in a solid: *pF* is the logarithm to the base 10 of the height in centimetres of a unit column of water.
pH: Abbreviation of 'potens hydrogen': a scale for measuring the acidity or alkalinity of mildly acid or alkaline substances, chosen so that the *pH* of water is 7, acid substances less than 7 and alkaline substances more than 7: *pH* is the negative logarithm to the base 10 of the hydrogen ion concentration.
PHOTOSYNTHESIS: The process whereby plants convert carbon dioxide into oxygen.
PICRITE: An ultra-basic igneous rock containing mainly augite and olivine.
PISOLITIC: Pea-like in structure: a structure consisting of cemented nodules or concretions.
PLAGIOCLASE: A group of feldspars containing sodium and calcium; they form a series from albite (soda plagioclase) to anorthite (calcic plagioclase) the latter weathering more readily than the former: see feldspar.
PLAYA: A lake-bed in arid areas: usually has a salt-encrusted surface and may fill with water after rain to form a playa lake.
PODZOL: A soil best developed under coniferous forests: it contains a surface mat of leaf litter over a bleached eluvial horizon: humus and iron are illuviated to the *B* horizons.
PODZOLISATION: Soil-forming processes occurring in cool, temperate regions, involving removal of bases, the accumulation of a surface mat of acid organic matter and illuviation of humus, iron and clay.
POLYGENESIS: The derivation of one soil profile from another as a result of a slow environmental change.
PRIMARY MINERAL: A mineral which has undergone neither deuteric alteration nor subaerial weathering.
PYROXENE: A group of readily weathered aluminosilicate minerals with single chain structures: the chains are linked mainly by calcium, magnesium and iron.

QUARTZ: One of the most common forms of crystalline silica (silicon oxide): very resistant to subaerial weathering.

RED EARTHS: A deep, loamy, red soil formed under free drainage in the humid tropics and dominated by sesquioxides.
REDUCTION: The addition of electrons to an atom or molecule, e.g., $Fe^{3+} + e \rightarrow Fe^{2+}$.
REG: Stony desert pavement.
REGOLITH: The whole mantle of weathering from the top of the soil down to fresh rock: it includes the solum and the soil parent material.
REGOSOL: An incipient soil formed on unconsolidated material such as alluvium.
RELICT SOIL: A soil whose morphology is not entirely explicable in terms of its present environment.
RENDZINA: A dark, calcareous, clayey soil found on limestone in sub-humid areas.
REVERSION OF PHOSPHATE: The change of soluble phosphate in superphosphate to insoluble phosphate in the soil.
RHIZOBIUM: The bacterium in the nodules of legumes which has the ability to fixate atmospheric nitrogen for use by the host plant.
RUTILE: Titanium oxide: TiO_2.

SAND: Mineral particles defined on the International (Atterberg) scale as having diameters between 2.0 and 0.02 mm.

SALT: Any combination of basic cation with acid anion, e.g., calcium carbonate (lime), sodium chloride (halite or common salt), magnesium sulphate and sodium bicarbonate.

SAVANNA, SAVANNAH: A climatic region within the tropics characterised by hot, wet summers and warm, dry winters: also used for the vegetation associated with that climate.

SCHIST: A low-grade, regionally metamorphosed rock with strong parallel structure and shiny appearance frequently containing a lot of mica and amphibole.

SECONDARY MINERAL: A mineral which has undergone some form of chemical change since its formation: the change may be subaerial or deuteric.

SEDIMENTARY ROCKS: Rocks which have formed by the consolidation of weathered particles deposited under water.

SELF-MULCHING SOIL: A soil with the natural ability to develop a thin horizon of crumb structure at the surface reducing water loss by free evaporation from the soil surface.

SERAL PROGRESSION: The progression of natural vegetation from the first established plant communities to the climatic climax.

SESQUIOXIDES: The hydrated oxides (hydroxides) of iron, aluminium, manganese and titanium: iron and aluminium oxides are the most common.

SILCRETE: Any material hardened by enrichment by silica.

SILICA: Silicon oxide—SiO_2.

SILICATE CLAYS: The secondary minerals formed by the break down of primary minerals: they are composed of silica and alumina sheets.

SILT: Mineral particles defined on the International (Atterberg) scale as having diameters between 0.02 and 0.002 mm.

SKELETAL SOIL: See Lithosol.

SOIL: Weathered mineral and organic matter found in genetically related horizons in response to subaerial processes.

SOIL ASSOCIATION: A group of soils commonly found together in the field, often in catenary relationship (see also Catena).

SOIL-FORMING FACTORS: The variables which control the soil-forming processes and soil morphology: there are five soil-forming factors—climate, relief, organisms (including vegetation), parent material and time or the duration of weathering.

SOIL-FORMING PROCESSES: The transformations within the soil which cause the development of horizons in response to the soil's environment: the main soil-forming processes are the movements out of or within the soil of bases, humus, iron, aluminium and silicate clay.

SOIL HORIZON: A layer in a soil formed by the action of subaerial forces on weathering material causing movements of constituents up and down the soil profile: soil horizons are genetically related to each other by the soil-forming processes.

SOIL MORPHOLOGY: The form or properties of the soil as seen in the field.

SOIL ORDER: The highest category in soil classification: often involves division into zonal, intrazonal and azonal soils in terms of the five factors of soil formation.

SOIL PARENT MATERIAL: The material from which the soil horizons have developed.

SOIL REACTION: Soil acidity or alkalinity, most conveniently measured on the *pH* scale (see *pH*).

SOIL SCIENCE: The study of all aspects of the soil, including soil physics, chemistry, pedology, microbiology, fertility, plant nutrition and mineralogy.

SOIL SUBORDER: The second level in soil classification, generally associated with the most obvious profile features.

SOLIFLUCTION: The movement of soil down slopes in cold regions whereby the soil, saturated with meltwater, moves sluggishly down the slope after the thaw.

SOLUM: The horizons of the soil which have been changed from the parent material by the soil-forming processes: the solum has a different organisation from the parent material.

SQUAMOSE: Soil structural aggregates which are platy or laminar.

STICKY POINT: See Field capacity. The moisture content at sticky point, when the soil just adheres to the fingers, approximates field capacity.

STOMATA: The holes in the leaves of plants through which plants transpire moisture (see Transpiration).

STRUCTURE: The arrangement of the soil: the form or orderliness of the soil mass: the natural units of soil structure are peds.

SUPERPHOSPHATE: The most common form of phosphate fertilizer used in Australia: $Ca\ (H_2PO_4)_2$.

SYMBIOTIC FIXATION OF NITROGEN: The taking in of nitrogen by the bacteria in the nodules of legumes for use by the host plant.

TERRA ROSSA: A red, acid, clayey soil found on limestone in humid areas.

TEXTURE: The rough field assessment of the soil's particle size distribution: the soil is assigned to a texture class after moulding a wet sample between the fingers.

TEXTURE TRIANGLE: A graphical method for plotting the results of mechanical analysis of soils: the texture triangle has sand, silt and clay at the vertices and the plotting axes are the sides of the triangle: the texture triangle is used to relate field texture classes to the results of mechanical analysis.

TILTH: The ability of a soil to form a finely granular structure suitable for sowing seeds.

TITANOMAGNETITE, TITANOMAGHAEMITE: Iron oxides in which some titanium substitutes for iron.

TITRATION: A procedure in chemistry whereby the amount of a chemical in solution is estimated by adding another solution of known concentration until a reaction occurs; the reaction may be observed by adding an indicator which changes colour itself at the reaction point.

TRACE ELEMENT: An element required by plants in minute quantities, but toxic at quite low concentrations.

TRANSPIRATION: Loss of water from the soil through the stomata of plants.

TRANSPIRATION RATIO OR COEFFICIENT: The units of water transpired by the plant in the production of every unit of above ground dry matter.

TRIPLE-SUPER: A widely-used phosphate fertilizer: $CaHPO_4$.

TROPOPHYTIC PLANTS: Plants which are hygrophytic in one season and xerophytic in another (see Hygrophytic and Xerophytic).

UNIFORM PROFILE: (Used in Northcote's system of soil classification.) A soil with little, if any, texture change down the profile.

VERMICULITE: A member of the hydrous mica group of clay minerals, with two silica and one alumina sheet forming the crystal lattice.

VESICULAR: A structure containing many voids.

XEROPHYTIC: Drought-resisting: xerophytic plants have mechanisms for surviving water shortage.

ZEOLITE: A group of readily weathered aluminosilicate minerals with a continuous framework of silicon and aluminium tetrahedra, linked by sodium, potassium and calcium.

ZIRCON: Zirconium oxide.

ZONAL SOILS: Soils which reflect mainly the effect of climate and the associated vegetation: zonal soil distribution broadly follows climatic regions.

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