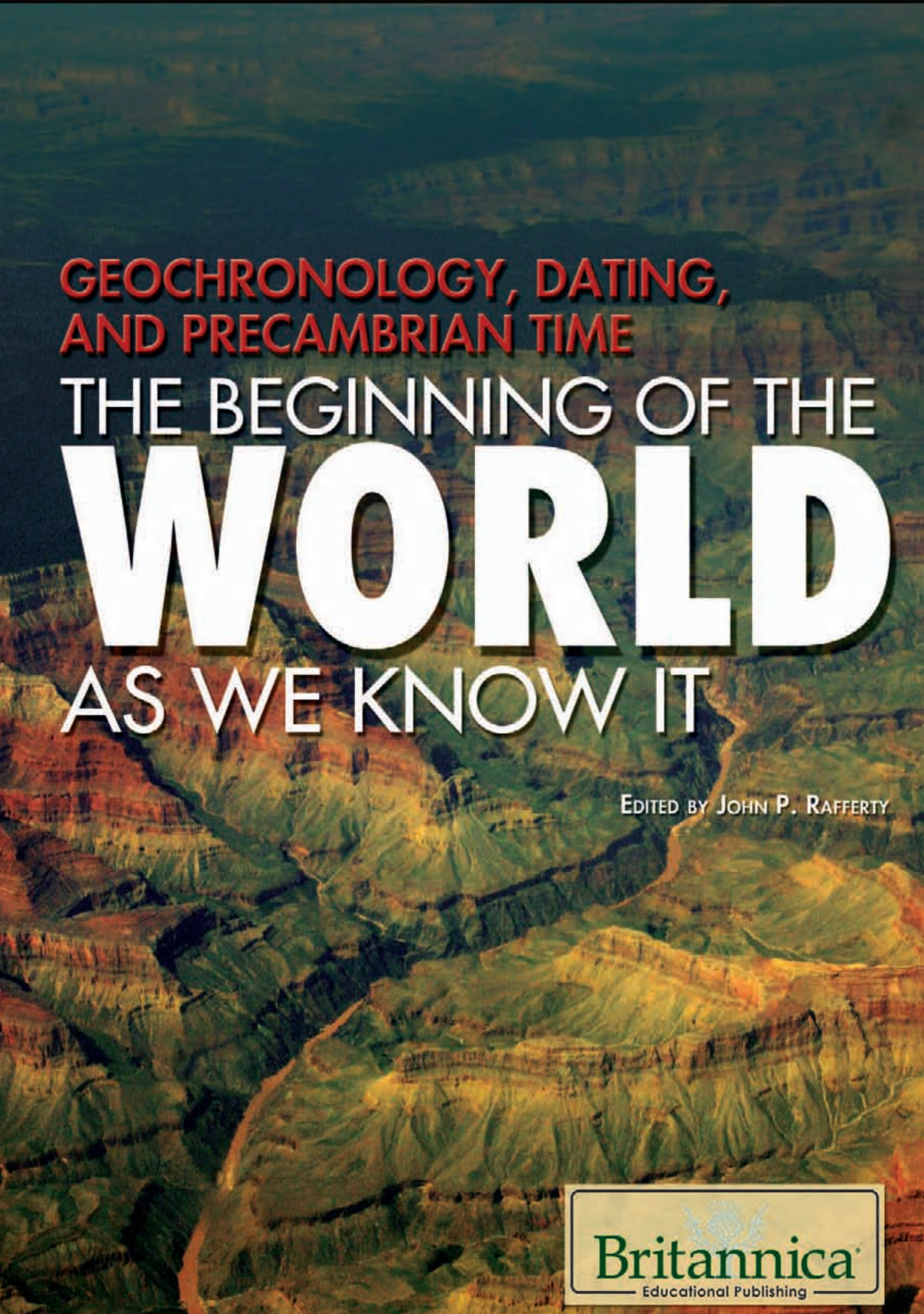
THE GEOLOGIC HISTORY OF EARTH



GEOCHRONOLOGY, DATING, AND PRECAMBRIAN TIME

THE BEGINNING OF THE WORLD AS WE KNOW IT

EDITED BY JOHN P. RAFFERTY

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EDITED BY JOHN P. RAFFERTY, ASSOCIATE EDITOR, EARTH SCIENCES

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On the cover: The Colorado River in the Grand Canyon, Arizona. *Shutterstock.com*

On page 12: This fossil is of *Dickinsonia*, a segmented flatworm that lived during the Precambrian era. *O. Louis Mazzatenta/National Geographic/Getty Images*

On pages 5, 23, 79, 156, 232, 233, 236, 238: The ripple marks in this Australian sandstone show the action of Precambrian oceans on soft, ancient rock. *Tim Graham Photo Library/Getty Images*

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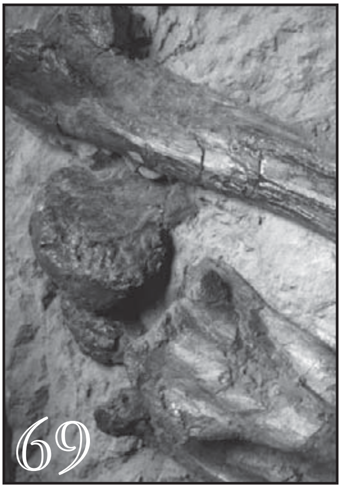
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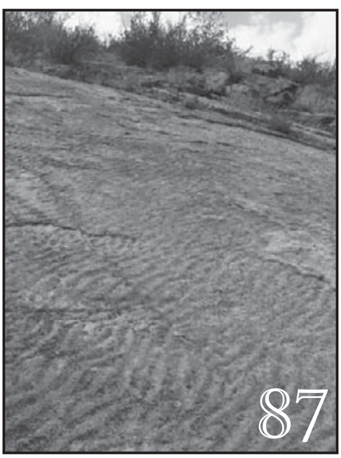
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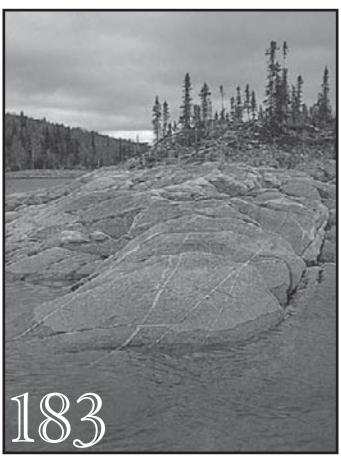
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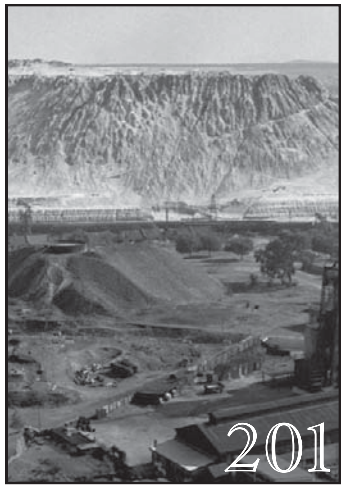
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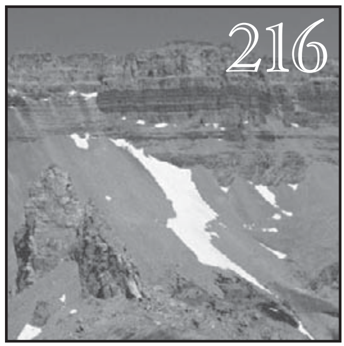
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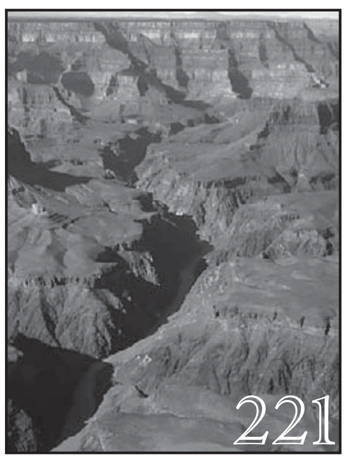
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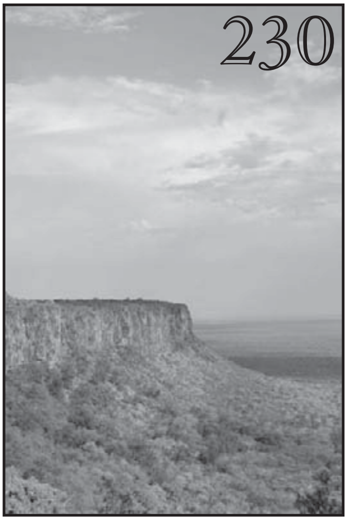
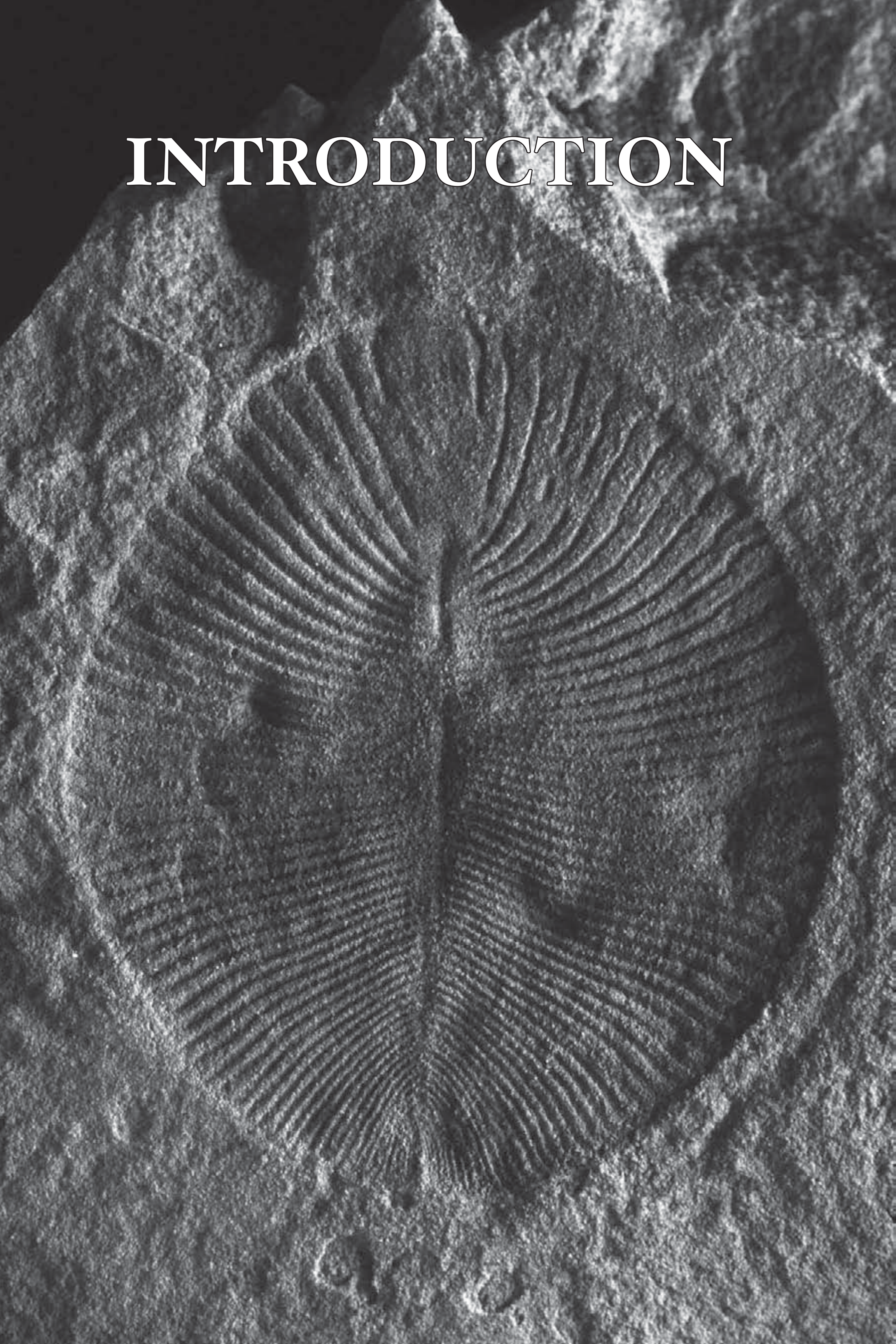


INTRODUCTION



Planet Earth was formed roughly 4.6 billion years ago. For human beings—used to measuring time in terms of days, weeks, and months—such an enormous span of time can be a difficult concept to grasp. Geologists, scientists who study the Earth and the processes that continue to shape it, have broken up this vast expanse of "deep time" into major divisions based on what they have learned from the study of ancient rocks and fossils. The first of these divisions—from approximately 4.6 billion until 542 million years ago—is known as the Precambrian, meaning everything that happened before the Cambrian period. (Today some people prefer to call this period the Cryptozoic, which means "hidden life.") Almost all of planet Earth's history is Precambrian. Until recently, however, it has remained the most unknown, the strangest, and most perplexing period in all geologic history—what some have referred to as the "Dark Ages" of Earth's existence.

For centuries, the Earth yielded no fossil record to help humans envision Precambrian time. Discoveries of rich caches of fossils from the Cambrian period enabled scientists to assemble a vivid picture of the creatures that inhabited the planet during that time, but the Precambrian Earth remained largely unimaginable. What was known as the "missing fossil record" of the Precambrian period stood for more than a century as one of the great unsolved mysteries of the natural sciences. This mystery perplexed Charles Darwin and many other scientists who followed in his footsteps. Humans had not yet developed the various methods to accurately determine the age of rocks formed during this interval of geologic time. They had not yet identified and interpreted the remains of the microscopic bacteria that formed in the earliest oceans. The story scientists were beginning to piece together was full of gaps and inconsistencies. It offered glimpses of an alien Earth with an unstable, roiling surface, rocked by volcanic events and cosmic collisions, alternated between extremes of ice and fire, and an atmosphere that would poison most life as we know it today. How did such a hellish place give rise—over the course of an almost unimaginable span of years—to all the familiar features of our planet: oceans, mountains, and valleys, and an oxygen-rich atmosphere that sustains the flowering of plant and animal life in all its countless forms?

Scientists who study the Precambrian period grapple with some of the most profound questions that human beings have ever asked: How old is the Earth? Where did the Moon come from? What made the oceans and the mountains and valleys? How and when did life begin? People of different cultures and religious faiths have shared creation stories to help explain these ancient mysteries. Paleogeologists and paleobiologists—scientists who study the ancient Earth and the life-forms that arose on it—have devised scientific tools to begin answering the same questions.

What has enabled scientists to speak with authority about what occurred on this planet billions of years ago? Geologists have developed systems of dating that enable them to study the geologic processes taking place on the planet today and make educated guesses about its past. James Hutton (1726–1797), considered by many to be the founder of modern geology, laid the groundwork for this earth science in 1785 when he presented his scientific papers at the Royal Society of Edinburgh in Scotland. Hutton's bold thesis had to do with the concept of geologic cycles—the recognition that processes such as erosion, deposition, sedimentation, and upthrusting are cyclical and must have been repeated many times over the long course of Earth's history, with "no vestige of a beginning, no prospect of an end." Entire mountain ranges rise up and are eroded away, over and over again. Hutton reasoned that since each geologic cycle takes many millennia to complete, the Earth must be far older than anyone had previously believed. The term "uniformitarianism" was introduced to describe this process by Cambridge scholar William Whewell in 1832.

The Scottish geologist Charles Lyell elaborated on Hutton's theory with his own theory of gradualism. When Charles Darwin embarked aboard the HMS *Beagle* on his legendary voyages, he brought along Charles Lyell's book *Principles of Geology* (1830). In this volume, Lyell sets forth the argument (radical at that time) that present-day geological processes can explain the history of the Earth. Many people believed that the biblical story of the flood, or some such cataclysmic event, accounted for Earth's geological features. Lyell instead argued that these features were produced gradually over millennia by geologic processes still occurring today, which have operated uniformly throughout history. (The motto of uniformitarian science might be summed up as "The present is the key to the past.")

Another benchmark in geochronology, the dating of events in the Earth's history, stemmed from William Smith's work with faunal sequence. During preparations for the digging of a coal canal in southwestern England in 1793, Smith observed that layers, or strata, of sedimentary rock contained fossils in a definite sequence, and that the same sequence could be found in rocks elsewhere. The discovery that fossil plants and animals succeed one another in time in a predictable manner, now known as the law of faunal succession, became one of the keys to unlocking the secrets of deep time. Another key was

Geologic Time Scale

Eonothem/ Eon	Erathem/ Era	System/ Period	Series/ Epoch	Stage/ Age	mya¹
Phanerozoic	Cenozoic	Quaternary	Holocene		0.0117
			Pleistocene	Tarantian	0.126
				"Ionian"	0.781
				Calabrian	1.806
				Gelasian	2.588
		Neogene	Pliocene	Piacenzian	3.600
				Zanclean	5.332
			Miocene	Messinian	7.246
				Tortonian	11.608
				Serravallian	13.82
				Langhian	15.97
				Burdigalian	20.43
				Aquitanian	23.03
		Paleogene	Oligocene	Chattian	28.4 ± 0.1
				Rupelian	33.9 ± 0.1
			Eocene	Priabonian	37.2 ± 0.1
				Bartonian	40.4 ± 0.2
				Lutetian	48.6 ± 0.2
				Ypresian	55.8 ± 0.2
			Paleocene	Thanetian	58.7 ± 0.2
				Selandian	~61.1
				Danian	65.5 ± 0.3
	Mesozoic	Cretaceous	Upper	Maastrichtian	70.6 ± 0.6
				Campanian	83.5 ± 0.7
				Santonian	85.8 ± 0.7
				Coniacian	~88.6
				Turonian	93.6 ± 0.8
				Cenomanian	99.6 ± 0.9
			Lower	Albian	112.0 ± 1.0
				Aptian	125.0 ± 1.0
				Barremian	130.0 ± 1.5
				Hauterivian	~133.9
				Valanginian	140.2 ± 3.0
				Berriasian	145.5 ± 4.0

Eonothem/ Eon	Erathem/ Era	System/ Period	Series/ Epoch		Stage/ Age	mya¹
Phanerozoic	Mesozoic	Jurassic	Upper		Tithonian	145.5 ± 4.0 150.8 ± 4.0
					Kimmeridgian	~155.6
					Oxfordian	161.2 ± 4.0
			Middle		Callovian	164.7 ± 4.0
					Bathonian	167.7 ± 3.5
					Bajocian	171.6 ± 3.0
					Aalenian	175.6 ± 2.0
			Lower		Toarcian	183.0 ± 1.5
					Pliensbachian	189.6 ± 1.5
					Sinemurian	196.5 ± 1.0
					Hettangian	199.6 ± 0.6
		Triassic	Upper		Rhaetian	203.6 ± 1.5
					Norian	216.5 ± 2.0
					Carnian	~228.7
			Middle		Ladinian	237.0 ± 2.0
					Anisian	~245.9
			Lower		Olenekian	~249.5
					Induan	251.0 ± 0.4
	Paleozoic	Permian	Lopingian		Changhsingian	253.8 ± 0.7
					Wuchiapingian	260.4 ± 0.7
			Guadalupian		Capitanian	265.8 ± 0.7
					Wordian	268.0 ± 0.7
					Roadian	270.6 ± 0.7
			Cisuralian		Kungurian	275.6 ± 0.7
					Artinskian	284.4 ± 0.7
					Sakmarian	294.6 ± 0.8
					Asselian	299.0 ± 0.8
		Carboniferous	Pennsylvanian²	Upper	Gzhelian	303.4 ± 0.9
					Kasimovian	307.2 ± 1.0
				Middle	Moscovian	311.7 ± 1.1
				Lower	Bashkirian	318.1 ± 1.3
			Mississippian²	Upper	Serpukhovian	328.3 ± 1.6
				Middle	Visean	345.3 ± 2.1
				Lower	Tournaisian	359.2 ± 2.5

¹ Millions of years ago.
² Both the Mississippian and Pennsylvanian time units are formally designated as sub-periods within the Carboniferous Period.
³ Several Cambrian unit age boundaries are informal and are awaiting ratified definitions.

Encyclopædia Britannica, Inc. Source: International Commission on Stratigraphy (ICS)

Eonothem/ Eon	Erathem/ Era	System/ Period	Series/ Epoch	Stage/ Age	mya[1]
Phanerozoic	Paleozoic	Devonian	Upper	Famennian	359.2 ± 2.5
				Frasnian	374.5 ± 2.6
			Middle	Givetian	385.3 ± 2.6
				Eifelian	391.8 ± 2.7
			Lower	Emsian	397.5 ± 2.7
				Pragian	407.0 ± 2.8
				Lochkovian	411.2 ± 2.8
		Silurian	Pridoli		416.0 ± 2.8
			Ludlow	Ludfordian	418.7 ± 2.7
				Gorstian	421.3 ± 2.6
			Wenlock	Homerian	422.9 ± 2.5
				Sheinwoodian	426.2 ± 2.4
			Llandovery	Telychian	428.2 ± 2.3
				Aeronian	436.0 ± 1.9
				Rhuddanian	439.0 ± 1.8
		Ordovician	Upper	Hirnantian	443.7 ± 1.5
				Katian	445.6 ± 1.5
				Sandbian	455.8 ± 1.6
			Middle	Darriwilian	460.9 ± 1.6
				Dapingian	468.1 ± 1.6
			Lower	Floian	471.8 ± 1.6
				Tremadocian	478.6 ± 1.7
		Cambrian[3]	Furongian	Stage 10	488.3 ± 1.7
				Stage 9	~492.0
				Paibian	~496.0
			Series 3	Guzhangian	~499.0
				Drumian	~503.0
				Stage 5	~506.5
			Series 2	Stage 4	~510.0
				Stage 3	~515.0
			Terreneuvian	Stage 2	~521.0
				Fortunian	~528.0
					542.0 ± 1.0

Eonothem/ Eon		Erathem/ Era	System/ Period	mya[1]
Precambrian	Proterozoic	Neoproterozoic	Ediacaran	542
			Cryogenian	~635
			Tonian	850
		Mesoproterozoic	Stenian	1,000
			Ectasian	1,200
			Calymmian	1,400
		Paleoproterozoic	Statherian	1,600
			Orosirian	1,800
			Rhyacian	2,050
			Siderian	2,300
	Archean	Neoarchean		2,500
		Mesoarchean		2,800
		Paleoarchean		3,200
		Eoarchean		3,600
	Hadean (informal)			4,000
				4,600

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Source: 2009 International Stratigraphic Chart produced by the ICS.

provided by French zoologist Georges Cuvier's 1812 hypothesis that fossils do record geologic events. These advances in geochronology were based on the phenomenon of stratification, the naturally occurring sequence of rocks in layers with the youngest rocks on the top and the oldest on the bottom. Stratigraphy, a branch of geology concerned with studying the stratification of rocks, enables scientists to determine the age of rocks and fossils relative to one another.

In the 20th century, scientists devised methods of dating rocks that were based on chemicals rather than fossil sequences. Antoine-Henri Becquerel's discovery of radioactivity in 1896 proved key to the development of radiometric dating techniques that revolutionized the science of geochronology. Some elements, such as uranium, undergo radioactive decay, a process that happens at a uniform and predictable rate. In 1905, John William Strutt was the first person to successfully apply a radiometric technique to the study of earth materials when he succeeded in determining the age of a radium-containing rock by analyzing its helium content.

Radioactive decay is a property of certain naturally occurring elements, and radioactive isotopes can also be created under laboratory conditions. Each radioactive isotope has a fixed rate of decay, called its half-life, during which radioactive "parent" atoms transform into "daughters," or atoms of a different chemical element. Naturally occurring uranium isotopes transform into lead, and rubidium isotopes become strontium as they decay. By determining the ratio of parent to daughter atoms, one can accurately calculate the age of a rock containing those elements.

Today the geologist's toolkit includes many radiometric techniques that use isotopes to determine the age of ancient rocks and fossils. There are two main methods of observing radioactive decay. The first of these detects and counts radioactive atoms through the radiation they emit. Carbon-14 dating is an example of this first method. This technique depends upon the fact that the radioactive isotope carbon-14 present in the atmosphere and contained in all organic matter will begin to decay once an organism has expired. By comparing the carbon-14 present in a dead object to the carbon-14 in living matter, one can determine when an organism died. The second method uses a device called a mass spectrometer to count the atoms of each isotope, sifting them one atom at a time according to weight. The potassium-argon dating method is an example of this second method. Unlike stratigraphic methods of dating that supply relative ages, radiometric methods enable scientists to determine the approximate age of an object. Both absolute and relative dating techniques are important and complementary tools in geochronology, however, since not every rock can be dated radiometrically.

Geologic processes can serve as absolute chronometers, as well. Measurements based on weathering processes (as in obsidian dating), tree-ring and coral growth, and variations in the Earth's magnetic field—where magnetic minerals in rocks provide the record of ancient change—have enabled geologists to precisely calculate the ages of rocks. The nature of the most ancient rocks suggests there are even older rocks that have yet to be found.

All these dating methods, along with recent fossil discoveries, have helped geologists formulate a clearer picture of the earliest period in Earth's history. Precambrian time is currently divided into three eons: the Hadean Eon, the Archean Eon, and the Proterozoic Eon. Each has a unique character and represents a major stage in the planet's life.

The Hadean (4.6 to about 4 billion years ago), which takes its name from Hades, the ancient Greek underworld, predates the formation of any rocks, so what people know about this time is largely based on computer models and guesswork. During the Hadean Eon, many scientists believe that the solar system formed out of gases and dust, the sun began to shine, and Earth took shape in the midst of meteor showers and other galactic debris.

The Archean Eon dates from approximately 4 billion to 2.5 billion years ago and marks the beginning of life and geological processes on Earth. During the Archean Eon, Earth's atmosphere was a brew of nitrogen, methane, and carbon dioxide. A crust formed on the surface of the cooling planet, and structures called stromatolites, consisting mainly of blue-green algae and other microorganisms, developed in the oceans. The oldest rocks on Earth date from this time, and all the planet's continents have Archean cores.

The Proterozoic Eon, the longest chapter in Earth's history, dates from 2.5 billion to 542 million years ago and includes the beginning of the geological process called plate tectonics. Moving plates on the planet's surface converged to create mountains and fractured entire continents when they pulled apart, a process that continues today. The oldest known glacial episode, or ice age, occurred in the Proterozoic Eon, during which all or nearly all of Earth's surface was covered by some amount of ice. During warmer periods, new varieties of bacteria began to harness the power of the sun through the biochemical process of photosynthesis. The oxygen they produced collected in the oceans and then in the atmosphere, sparking an ecological event known as the Great Oxidation Event, creating conditions poisonous to the anaerobic life-forms on Earth. The ancient anaerobes retreated underground, and oxygen-dependent life-forms began to evolve.

Studying the Precambrian is like having a front-row seat at the creation of the Earth. Here, in deep time, one finds the oldest minerals and rocks, the earliest oceans, and the first stirrings of biological life—evidence of which has been found in western Greenland. The Precambrian period also contained the first stages of sexual differentiation and the beginnings of sexual reproduction with the appearance of the first eukaryotes, a huge group of organisms that would eventually include protozoans, fungi, plants, animals, and, finally, human beings.

Contemporary concerns about global warming have spurred interest in paleoclimatology, the study of climatic conditions of past geologic ages. Scientists know that during the long course of Precambrian time, the climatic conditions of Earth changed radically. Evidence of this can be found in the sedimentary record, which documents significant changes in the composition of the atmosphere and oceans over time. The Earth's deepest secrets still abide in the stones under our feet, but the more people learn about the planet's Precambrian history, the better prepared we may be to face the challenges of the present. Geological science teaches us that the present is the key to the past, but we may find, before long, that this knowledge is also the key to the future.

CHAPTER 1

Geologic Time

Geologic time spans Earth's geologic history. It extends from about 4.6 billion years ago (corresponding to the age of Earth's formation) to the present day. Some scientists maintain, however, that geologic time begins with the oldest known rocks that were created some 4.4 billion years ago.

The geologic time scale is the "calendar" for events in Earth history. It subdivides all time since the end of the Earth's formative period as a planet (nearly 4.6 billion years ago) into named units of abstract time: the latter, in descending order of duration, are eons, eras, periods, and epochs. The enumeration of these geologic time units is based on stratigraphy, which is the correlation and classification of rock strata. The fossil forms that occur in these rocks provide the chief means of establishing a geologic time scale. Because living things have undergone evolutionary changes over geologic time, particular kinds of organisms are characteristic of particular parts of the geologic record. By correlating the strata in which certain types of fossils are found, the geologic history of various regions (and of the Earth as a whole) can be reconstructed. The relative geologic time scale developed from the fossil record has been numerically quantified by means of absolute dates obtained with radiometric dating methods.

GEOCHRONOLOGY

Geochronology is the field of scientific investigation concerned with determining the age and history of the Earth's rocks and rock assemblages. Such time determinations are made and the record of past geologic events is deciphered by studying the distribution and succession of rock strata, as well as the character of the fossil organisms preserved within the strata.

The Earth's surface is a complex mosaic of exposures of different rock types that are assembled in an astonishing array of geometries and sequences. Individual rocks in the myriad of rock outcroppings (or in some instances shallow subsurface occurrences) contain certain materials or mineralogic information that can provide insight as to their "age."

For years investigators determined the relative ages of sedimentary rock strata on the basis of their positions in an outcrop and their fossil content. According to a long-standing principle of the geosciences, that of superposition, the oldest layer within a sequence of strata is at the base and the layers are progressively younger with ascending order. The relative ages of the rock strata deduced in this manner can be corroborated and at times refined by the examination of the fossil forms present. The tracing and matching of the fossil content of separate rock outcrops (that is, the correlation process) eventually enabled investigators to integrate rock sequences in many areas of the world and construct a relative geologic time scale.

Scientific knowledge of the Earth's geologic history has advanced significantly since the development of radiometric dating, a method of age determination based on the principle that radioactive atoms in geologic materials decay at constant, known rates to daughter atoms. Radiometric dating has provided not only a means of numerically quantifying geologic time but also a tool for determining the age of various rocks that predate the appearance of life-forms.

EARLY VIEWS AND DISCOVERIES

Some estimates suggest that as much as 70 percent of all rocks outcropping from the Earth's surface are sedimentary. Preserved in these rocks is the complex record of the many transgressions and regressions of the sea, as well as the fossil remains or other indications of now extinct organisms and the petrified sands and gravels of ancient beaches, sand dunes, and rivers.

This ancient dragonfly was fossilized in limestone. Many species of dragonflies lived during the Precambrian era. Colin Keates/Dorling Kindersley/Getty Images

Modern scientific understanding of the complicated story told by the rock record is rooted in the long history of observations and interpretations of natural phenomena extending back to the early Greek scholars. Xenophanes of Colophon (560?–478? BCE), for one, saw no difficulty in describing the various seashells and images of life-forms embedded in rocks as the remains of long-deceased organisms. In the correct spirit but for the wrong reasons, Herodotus (5th century BCE) felt that the small discoidal nummulitic petrifactions (actually the fossils of ancient lime-secreting marine protozoans) found in limestones outcropping at al-Jīzah, Egypt, were the preserved remains of discarded lentils left behind by the builders of the pyramids.

These early observations and interpretations represent the unstated origins of what was later to become a basic principle of uniformitarianism, the root of any attempt at linking the past (as preserved in the rock record) to the present. Loosely stated, the principle says that the various natural phenomena observed today must also have existed in the past.

Although quite varied opinions about the history and origins of life and of the Earth itself existed in the pre-Christian era, a divergence between Western and Eastern thought on the subject of natural history became more pronounced as a result of the extension of Christian dogma to the explanation of natural phenomena. Increasing constraints were placed upon the interpretation of nature in view of the teachings of the Bible. This required that the Earth be conceived of as a static, unchanging body, with a history that began in the not too distant past, perhaps as little as 6,000 years earlier. It also required an end, according to the scriptures, that was in the not too distant future. This biblical history of the Earth left little room for interpreting the Earth as a dynamic, changing system. Past catastrophes, particularly those that may have been responsible for altering the Earth's surface such as the great flood of Noah, were considered an artifact of the earliest formative history of the Earth. As such, they were considered unlikely to recur on what was thought to be an unchanging world.

With the exception of a few prescient individuals such as Roger Bacon (*c.* 1220–92) and Leonardo da Vinci (1452–1519), no one stepped forward to champion an enlightened view of the natural history of the Earth until the mid-17th century. Leonardo seems to have been among the first of the Renaissance scholars to "rediscover" the uniformitarian dogma through his observations of fossil marine organisms and sediments exposed in the hills of northern Italy. He recognized that the marine organisms now found as fossils in rocks exposed in the Tuscan Hills were simply ancient animals that lived in the region when it had been covered by the sea and were eventually buried by muds along the seafloor. He also recognized that the rivers of northern Italy, flowing south from the Alps and emptying into the sea, had done so for a very long time.

In spite of this deductive approach to interpreting natural events and the possibility that they might be preserved and later observed as part of a rock outcropping, little or no attention was given to the history—namely, the sequence of events in their natural progression—that might be preserved in these same rocks.

The Principle of Superposition of Rock Strata

In 1669 the Danish-born natural scientist Nicolaus Steno (né Niels Steensen) published his noted treatise *De solido intra solidum naturaliter contento dissertationis prodromus* (Eng. trans. *The Prodromus of Nicolaus Steno's Dissertation*

Concerning a Solid Body Enclosed by Process of Nature Within a Solid). This seminal work laid the essential framework for the science of geology by showing in very simple fashion that the layered rocks of Tuscany exhibit sequential change—that they contain a record of past events. Following from this observation, Steno concluded that the Tuscan rocks demonstrated superpositional relationships: rocks deposited first lie at the bottom of a sequence, while those deposited later are at the top. This is the crux of what is now known as the principle of superposition. Steno put forth still another idea—that layered rocks were likely to be deposited horizontally. Therefore, even though the strata of Tuscany were (and still are) displayed in anything but simple geometries, Steno's elucidation of these fundamental principles relating to the formation of stratified rock made it possible to work out not only superpositional relationships within rock sequences but also the relative age of each layer.

With the publication of the *Prodromus* and the ensuing widespread dissemination of Steno's ideas, other natural scientists of the latter part of the 17th and early 18th centuries applied them to their own work. The early English geologist John Strachey, for example, produced in 1725 what may well have been the first modern geologic maps of rock strata. He also described the succession of strata associated with coal-bearing sedimentary rocks in Somersetshire, the same region of England where he had mapped the rock exposures.

The Classification of Stratified Rocks

In 1756 Johann Gottlob Lehmann of Germany reported on the succession of rocks in the southern part of his country and the Alps, measuring and describing their compositional and spatial variation. While making use of Steno's principle of superposition, Lehmann recognized the existence of three distinct rock assemblages: (1) a successionally lowest category, the Primary (Urgebirge), composed mainly of crystalline rocks, (2) an intermediate category, or the Secondary (Flötzgebirge), composed of layered or stratified rocks containing fossils, and (3) a final or successionally youngest sequence of alluvial and related unconsolidated sediments (Angeschwemmtgebirge) thought to represent the most recent record of the Earth's history.

German geologist Johann Gottlob Lehmann. Apic/Hulton Archive/ Getty Images

This threefold classification scheme was successfully applied with minor alterations to studies in other areas of Europe by three of Lehmann's contemporaries. In Italy, again in the Tuscan Hills in the vicinity of Florence, Giovanni Arduino—regarded by many as the father of Italian geology—proposed a four-component rock succession. His Primary and Secondary divisions are roughly similar to Lehmann's Primary and Secondary categories. In addition, Arduino proposed another category, the Tertiary division, to account for poorly consolidated though stratified fossil-bearing rocks that were superpositionally older than the (overlying) alluvium but distinct and separate from the hard (underlying) stratified rocks of the Secondary.

In two separate publications, one that appeared in 1762 and the second in 1773, Georg Christian Füchsel also applied Lehmann's earlier concepts of superposition to another sequence of stratified rocks in southern Germany. While using upwards of nine separate categories of sedimentary rocks, Füchsel essentially identified discrete rock bodies of unique composition, lateral extent, and position within a rock succession. (These rock bodies would constitute formations in modern terminology.)

Nearly 1,000 kilometres (620 miles) to the east, the German naturalist Peter Simon Pallas was studying rock sequences exposed in the southern Urals of eastern Russia. His 1777 report differentiated a threefold division of rock, essentially reiterating Lehmann's work by extension.

Thus, by the latter part of the 18th century, the superpositional concept of rock strata had been firmly established through a number of independent investigations throughout Europe. Although Steno's principles were being widely applied, there remained to be answered a number of fundamental questions relating to the temporal and lateral relationships that seemed to exist among these disparate European sites. Were these various German, Italian, and Russian sites at which Lehmann's threefold rock succession was recognized contemporary? Did they record the same series of geologic events in the Earth's past? Were the various layers at each site similar to those of other sites? In short, was correlation among these various sites now possible?

THE EMERGENCE OF MODERN GEOLOGIC THOUGHT

Inherent in many of the assumptions underlying the early attempts at interpreting natural phenomena in the latter part of the 18th century was the ongoing controversy between the biblical view of Earth processes and history and a more direct approach based on what could be observed and understood from various physical relationships demonstrable in nature. A substantial amount of information about the compositional character of many rock sequences was beginning to accumulate at this time. Abraham Gottlob Werner, a scholar of wide repute and following from the School of Mining in Freiberg, Ger., was very successful in reaching a compromise between what could be said to be scientific "observation" and biblical "fact." Werner's theory was that all rocks (including the sequences being identified in various parts of Europe at that time) and the Earth's topography were the direct result of either of two processes: (1) deposition in the primeval ocean, represented by the Noachian flood (his two "Universal," or Primary, rock series), or (2) sculpturing and deposition during the retreat of this ocean from the land (his two "Partial," or disintegrated, rock series). Werner's interpretation, which came to represent the so-called Neptunist conception of the Earth's beginnings, found widespread and nearly universal acceptance owing in large part to its theological appeal and to Werner's own personal charisma.

One result of Werner's approach to rock classification was that each unique lithology in a succession implied its own unique time of formation during the Noachian flood and a universal distribution. As more and more comparisons were made of diverse rock outcroppings, it began to become apparent that Werner's interpretation did not "universally" apply. Thus, an increasingly vocal challenge to the Neptunist theory arose.

James Hutton's Recognition of the Geologic Cycle

In the late 1780s, the Scottish scientist James Hutton launched an attack on much of the geologic dogma that had its basis in either Werner's Neptunist approach or its corollary that the prevailing configuration of the Earth's surface is largely the result of past catastrophic events which have no modern counterparts. Perhaps the quintessential spokesman for the application of the scientific method in solving problems presented in the complex world of natural history, Hutton took issue with the catastrophist and Neptunist approach to interpreting rock histories and instead used deductive reasoning to explain what he saw. By Hutton's account, the Earth could not be viewed as a simple, static world not currently undergoing change. Ample evidence from Hutton's Scotland provided the key to unraveling the often thought but still rarely stated premise that events occurring today at the Earth's surface—namely erosion, transportation and deposition of sediments, and volcanism—seem to have their counterparts preserved in the rocks. The rocks of the Scottish coast and the area around Edinburgh proved the catalyst for his argument that the Earth is indeed a dynamic, ever-changing system, subject to a sequence of recurrent cycles of erosion and deposition and of subsidence and uplift. Hutton's formulation of the principle of uniformitarianism—which holds that Earth processes occurring today had their counterparts in the ancient past, while not the first time that this general concept was articulated—was probably the most important geologic concept developed out of rational scientific thought of the 18th century. The publication of Hutton's two-volume *Theory of the Earth* in 1795 firmly established him as one of the founders of modern geologic thought.

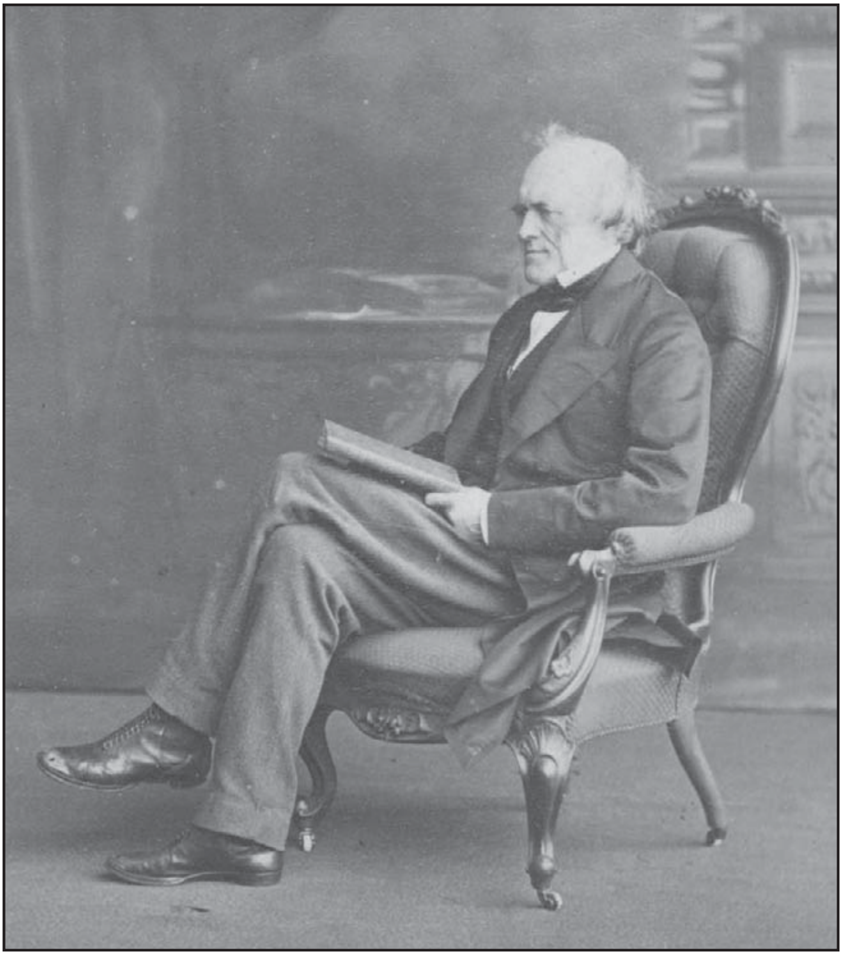
It was not easy for Hutton to popularize his ideas, however. The *Theory of the Earth* certainly did set the fundamental principles of geology on a firm basis, and several of Hutton's colleagues, notably John Playfair with his *Illustrations of the Huttonian Theory of the Earth* (1802), attempted to counter the entrenched Wernerian influence of the time. Nonetheless, another 30 years were to pass before Neptunist and catastrophist views of Earth history were finally replaced by those grounded in a uniformitarian approach.

This gradual unseating of the Neptunist theory resulted from the accumulated evidence that increasingly called into question the applicability of Werner's Universal and Partial formations in describing various rock successions. Clearly, not all assignable rock types would fit into Werner's categories, either superpositionally in some local succession or as a unique occurrence at a given site. Also, it was becoming increasingly difficult to accept certain assertions of Werner that some rock types (such as basalt) are chemical precipitates from the primordial ocean. It was this latter observation that finally rendered the Neptunist theory unsustainable. Hutton observed that basaltic rocks exposed in the Salisbury Crags, just on the outskirts of Edinburgh, seemed to have baked adjacent enclosing sediments lying both below and above the basalt. This simple observation indicated that the basalt was emplaced within the sedimentary succession while it was still sufficiently hot to have altered the sedimentary material. Clearly, basalt could not form in this way as a precipitate from the primordial ocean as Werner had claimed. Furthermore, the observations at Edinburgh indicated that the basalt intruded the sediments from below—in short, it came from the Earth's interior, a process in clear conflict with Neptunist theory.

While explaining that basalt may be intrusive, the Salisbury Craigs observations did not fully satisfy the argument that some basalts are not intrusive. Perhaps the Neptunist approach had some validity? The resolution of this latter problem occurred at an area of recent volcanism in the Auvergne area of central France. Here, numerous cinder cones and fresh lava flows composed of basalt provided ample evidence that this rock type is the solidified remnant of material ejected from the Earth's interior, not a precipitate from the primordial ocean.

Lyell's Promulgation of Uniformitarianism

Hutton's words were not lost on the entire scientific community. Charles Lyell, another Scottish geologist, was a principal proponent of Hutton's approach, emphasizing gradual change by means of known geologic processes. In his own observations on rock and faunal successions, Lyell was able to demonstrate the validity of Hutton's doctrine of uniformitarianism and its importance as one of the fundamental philosophies of the geologic sciences. Lyell, however, imposed some conditions on uniformitarianism that perhaps had not been intended by Hutton. He took a literal approach to interpreting the principle of uniformity in nature by assuming that all past events must have



Scottish geologist Sir Charles Lyell, Baronet. Ernest Edwards/Hulton Archive/Getty Images

conformed to controls exerted by processes that behaved in the same manner as those processes behave today. No accommodation was made for past conditions that do not have modern counterparts. In short, volcanic eruptions, earthquakes, and other violent geologic events may indeed have occurred earlier in Earth history but no more frequently nor with greater intensity than today. Accordingly, the surface features of the Earth are altered very gradually by a series of small changes rather than by occasional cataclysmic phenomena.

Lyell's contribution enabled the doctrine of uniformitarianism to finally hold sway, even though it did impose for the time being a somewhat limiting condition on the uniformity principle. This, along with the increased recognition of the utility of fossils in interpreting rock successions, made it possible to begin addressing the question of the meaning of time in Earth history.

Determining the Relationships of Fossils with Rock Strata

During this period of confrontation between the proponents of Neptunism and uniformitarianism, there emerged evidence resulting from a lengthy and detailed study of the fossiliferous strata of the Paris Basin that rock successions were not necessarily complete records of past geologic events. In fact, significant breaks frequently occur in the superpositional record. These breaks affect not only the lithologic character of the succession but also the character of the fossils found in the various strata.

An 1812 study by the French zoologist Georges Cuvier was prescient in its recognition that fossils do in fact record events in Earth history and serve as more than just "follies" of nature. Cuvier's thesis, based on his analysis of the marine invertebrate and terrestrial vertebrate fauna of the Paris Basin, showed conclusively that many fossils, particularly those of terrestrial vertebrates, had no living counterparts. Indeed, they seemed to represent extinct forms, which, when viewed in the context of the succession of strata with which they were associated, constituted part of a record of biological succession punctuated by numerous extinctions. These, in turn, were followed by a seeming renewal of more advanced but related forms and were separated from each other by breaks in the associated rock record. Many of these breaks were characterized by coarser, even conglomeratic strata following a break, suggesting "catastrophic" events that may have contributed to the extinction of the biota. Whatever the actual cause, Cuvier felt that the evidence provided by the record of faunal succession in the Paris Basin could be interpreted by invoking recurring catastrophic geologic events, which in turn contributed to recurring massive faunal extinction, followed at a later time by biological renewal.

As Cuvier's theory of faunal succession was being considered, William Smith, a civil engineer from the south of England, was also coming to realize that certain fossils can be found consistently associated with certain strata. In the course of evaluating various natural rock outcroppings, quarries, canals, and mines during the early 1790s, Smith increasingly utilized the fossil content as well as the lithologic character of various rock strata to identify the successional position of different rocks, and he made use of this information to effect a correlation among various localities he had studied. The consistency of the relationships that Smith observed eventually led him to conclude that there is indeed faunal succession and that there appears to be a consistent progression of forms from more primitive to more advanced. As a result of this observation, Smith was able to begin what was to amount to a monumental effort at synthesizing all that was then known of the rock successions outcropping throughout parts of Great Britain. This effort culminated in the publication of his "Geologic Map of England, Wales and Part of Scotland" (1815), a rigorous treatment of diverse geologic information resulting from a thorough understanding of geologic principles, including those of original horizontality, superposition (lithologic, or rock, succession), and faunal succession. With this, it now became possible to assume within a reasonable degree of certainty that correlation could be made between and among widely separated areas. It also became apparent that many sites that had previously been classified according to the then-traditional views of Arduino, Füchsel, and Lehmann did not conform to the new successional concepts of Smith.

Early Attempts at Mapping and Correlation

The seminal work of Smith at clarifying various relationships in the interpretation of rock successions and their correlations elsewhere resulted in an intensive look at what the rock record and, in particular, what the fossil record had to say about past events in the long history of the Earth. A testimony to Smith's efforts in producing one of the first large-scale geologic maps of a region is its essential accuracy in portraying what is now known to be the geologic succession for the particular area of Britain covered.

The application of the ideas of Lyell, Smith, Hutton, and others led to the recognition of lithologic and paleontologic successions of similar character from widely scattered areas. It also gave rise to the realization that many of these similar sequences could be correlated.

The French biologist Jean-Baptiste de Monet, Chevalier de Lamarck, in particular, was able to demonstrate the similarity of fauna from a number of Cuvier's and Alexandre Brongniart's collections of fossils from the Paris Basin with fossil fauna from the sub-Apennines of Italy and the London Basin. While based mainly on the collections of Cuvier and Brongniart, Lamarck's observations provided much more insight into the real significance of using fossils strictly for correlation purposes. Lamarck disagreed with Cuvier's interpretation of the meaning of faunal extinction and regeneration in stratigraphic successions. Not convinced that catastrophes caused massive and widespread disruption of the biota, Lamarck preferred to think of organisms and their distribution in time and space as responding to the distribution of favourable habitats. If confronted with the need to adapt to abrupt changes in local habitat—Cuvier's catastrophes—faunas must be able to change in order to survive. If not, they became extinct. Lamarck's approach, much like that of Hutton, stressed the continuity of processes and the continuum of the stratigraphic record. Moreover, his view that organisms respond to the conditions of their environment had important implications for the uniformitarian approach to interpreting Earth history.

Once it was recognized that many of the rocks of the Paris Basin, London Basin, and parts of the Apennines apparently belonged to the same sequence by virtue of the similarity of their fossil content, Arduino's term Tertiary (proposed as part of his fourfold division of rock succession in the Tuscan Hills of Italy) began to be applied to all of these diverse locations. Further work by Lyell and Gérard-Paul Deshayes resulted in the term Tertiary being accepted as one of the fundamental divisions of geologic time.

The Concepts of Facies, Stages, and Zones

Facies, stages, and zones are useful labels for describing and organizing geologic information. Facies is a general term used to describe lateral variations in rock that result from changes in species assemblages, rock types, and the effects of physical forces altering the rock outcrop. In contrast, stages and zones are typically vertical features. Stages indicate unique intervals of time within the rock. They are often represented by a sequential pattern of fossils and rock types. The boundaries between stages are often characterized by transitions in rock type and faunal composition. Zones are subdivisions of stages.

Facies

During the latter half of the 18th and early 19th centuries, most of the research on the distribution of rock strata and their fossil content treated lithologic boundaries as events in time representing limits to strata that contain unique lithology and perhaps a unique fossil fauna, all of which are the result of unique geologic processes acting over a relatively brief period of time. Hutton recognized early on, however, that some variations occur in the sediments and fossils of a given stratigraphic unit and that such variations might be related to differences in depositional environments. He noted that processes such as erosion in the mountains of Scotland, transportation of sand and gravels in streams flowing from these mountains, and the deposition of these sediments could all be observed to be occurring concurrently. At a given time then, these diverse processes were all taking place at separate locations. As a consequence, different environments produce different sedimentary products and may harbour different organisms. This aspect of differing lithologic type or environmental or biological condition came to be known as facies. (It was Steno who had, in 1669, first used the term facies in reference to the condition or character of the Earth's surface at a particular time.)

The significance of the facies concept for the analysis of geologic history became fully apparent with the findings of the Swiss geologist Amanz Gressly. While conducting survey work in the Jura Mountains in 1838, Gressly observed that rocks from a given position in a local stratigraphic succession frequently changed character as he traced them laterally. He attributed this lateral variation to lateral changes in the depositional environments responsible for producing the strata in question. Having no term to apply to the observed changes, he adopted the word "facies."

While Gressly employed the term specifically in the context of lithologic character, it is applied more broadly today. As now used, the facies concept has come to encompass other types of variation that may be encountered as one moves laterally (such as along outcroppings of rock strata exposed in stream valleys or mountain ridges) in a given rock succession. Lithologic facies, biological facies, and even environmental facies can be used to describe sequences of rocks of the same or different age having a particularly unique character.

In 1838, geologist Amanz Gressly studied lateral variations in ancient rock strata in Switzerland's Jura Mountains. Shutterstock.com

Stages and Zones

The extensive review of the marine invertebrate fauna of the Paris Basin by Deshayes and Lyell not only made possible the formalization of the term Tertiary but also had a more far-reaching effect. The thousands of marine invertebrate fossils studied by Deshayes enabled Lyell to develop a number of subdivisions of the Tertiary of the Paris Basin based on the quantification of molluskan species count and duration. Lyell noted that of the various assemblages of marine mollusks found, those from rocks at the top of the succession contained a large number of species that were still extant in modern environments. Progressively older strata yielded fewer and fewer forms that had living counterparts, until at the base of the succession, a very small number of the total species present could be recognized as having modern counterparts. This fact allowed Lyell to consider subdividing the Tertiary of the Paris Basin into smaller increments, each of which could be defined according to some relative percentage of living species present in the strata. The subdivision resulted in the delineation of the Eocene, Miocene, and Pliocene epochs in 1833. Later, this scheme was refined to further divide the Pliocene into an Early and a Late Pliocene.

Lyell's biostratigraphically defined concept of sequence, firmly rooted in concepts of faunal succession and superposition, was developed on mixed but stratigraphically controlled collections of fossils. It worked, but it did not address the faunal composition of the various Paris Basin strata other than in gross intervals—intervals that were as much lithologically as paleontologically defined. Alcide d'Orbigny, a French geologist, demonstrated correlational and superpositional uniqueness by utilizing paleontologically distinct intervals of strata defined solely on the basis of their fossil assemblages in his study of the French Jurassic *Terrains Jurassiques* (1842). This departure from a lithologically based concept of paleontologic succession enabled d'Orbigny to define paleontologically unique stages. Each stage represented a unique period in time and formed the basis of later work that resulted in the further subdivision of d'Orbigny's original stages into 10 distinct stage assemblages. In spite of the work of Smith and to a lesser extent Lyell and others, d'Orbigny's approach was essentially that of a catastrophist. Stage boundaries were construed to represent unusual extrinsic geologic events, with significant implications for faunal continuity. The applicability of d'Orbigny's stages to areas outside of France had only limited success. At this point in the development of paleontology as a science, little was understood about the geologic time range of various fauna. Even less was known about the habitats—the environmental limits—of ancient fauna. Could certain groups of organisms have sufficiently widespread distribution in the rock record to enable correlations to be made with certainty? The Jurassic of western Europe consisted mostly of shallow marine sediments widely deposited throughout the area. It is now known that some of the mollusks with which d'Orbigny worked were undergoing very rapid evolutionary change. They were thus relatively short-lived as distinct forms in the geologic record and had a wide-ranging environmental tolerance. The result was that some forms, notably of the group of mollusks called ammonite cephalopods, were distributed extensively within a variety of sedimentary facies. The correlating of strata based on the faunal stage approach was widely accepted. Interestingly, most of d'Orbigny's Jurassic stages, with refinements, are still in use today.

Only a short time after d'Orbigny's original analysis of Jurassic strata, the German mineralogist and paleontologist Friedrich A. Quenstedt challenged (in 1856–58) the validity of using stages to effect correlations in cases where the actual geologic ranges and bed-by-bed distribution of individual component fossils of an assemblage were unknown. In retrospect, this seems blatantly obvious, but at the time the systematic stratigraphic documentation of fossil occurrence was not always carried out. Much critical biostratigraphic data necessary for the proper characterization of faunal assemblages was simply not collected. As argued, individual fossil ranges and their distributions could have profound influence on the concept of faunal succession and evolutionary dynamics.

Several of Quenstedt's students at the University of Tübingen followed up on this latter concern. One in particular, Carl Albert Oppel, essentially refined his mentor's concepts by paying particular attention to the character of the range of individual species in a succession of fauna. These intervals of unique biological character, which he called zones, were essentially subdivisions of the stages proposed by Quenstedt. Oppel's recognition of the earliest occurrence of a fossil species (or its first appearance), its range through a succession of strata, and its eventual loss from the local record (or its last appearance) led him to compare such biostratigraphic data from many species. By making use of such data on species that overlap in some or all of their stratigraphic ranges and from widely separated areas, Oppel was able to erect a biochronology based on a diverse record of first appearances, last appearances, and individual and overlapping range zones. This fine-scale refinement of a biologically defined sense of succession found wide applicability and enabled not only biochronological (or temporal) but also biofacies (spatial) understanding of the succession in question.

THE COMPLETION OF THE PHANEROZOIC TIME SCALE

With the development of the basic principles of faunal succession and correlation and the recognition of facies variability, it was a relatively short step before large areas of Europe began to be placed in the context of a global geologic succession. This was not, however, accomplished in a systematic manner. Whereas the historical ideas of Lehmann and Arduino were generally accepted, it became increasingly clear that many diverse locally defined rock successions existed, each with its own unique fauna and apparent position within some sort of "universal" succession.

As discussed above, Arduino's Tertiary was recognized in certain areas and was in fairly common use after 1760, but only rudimentary knowledge of other rock successions existed by the later part of the 18th century. The German naturalist Alexander von Humboldt had recognized the widespread occurrence of fossil-bearing limestones throughout Europe. Particular to these limestones, which formed large tracts of the Jura Mountains of Switzerland, were certain fossils that closely resembled those known from the Lias and Oolite formations of England, which were then being described by William Smith. Subsequently, Humboldt's "Jura Kalkstein" succession, as he described it in 1795, came to be recognized throughout Europe and England. By 1839, when the geologist Leopold Buch recognized this rock sequence in southern Germany, the conceptual development of the Jurassic System was complete.

The coal-bearing strata of England, known as the Coal Measures, had been exploited for centuries, and their distribution and vertical and lateral variability were the subject of numerous local studies throughout the 17th and early 18th centuries, including those of Smith. In 1808 the geologist Jean-Baptiste-Julien d'Omalius d'Halloy described a coal-bearing sequence in Belgium as belonging to the Terrain Bituminifére. Although the name did not remain in common usage for long, the Terrain Bituminifére found analogous application in the work of two English geologists, William D. Conybeare and William Phillips, in their synthesis of the geology of England and Wales in 1822. Conybeare and Phillips coined the term Carboniferous (or coal-bearing) to apply to the succession of rocks from north-central England that contained the Coal Measures. The unit also included several underlying rock formations extending down into what investigators now consider part of the underlying Devonian System. At the time, however, the approach by Conybeare and Phillips was to encompass in their definition of the Carboniferous all of the associated strata that could be reasonably included in the Coal Measures succession.

D'Omalius mapped and described a local succession in western France. While doing so, he began to recognize a common sequence of soft limestones, greensands (glauconite-bearing sandstones), and related marls in what is today known to be a widespread distribution along coastal regions bordering the North Sea and certain regions of the Baltic. The dominant lithology of this sequence is frequently the soft limestones or chalk beds so well known from the Dover region of southeast England and Calais in nearby France. D'Omalius called this marl, greensand, and chalk-bearing interval the Terrain Crétacé. Along with their adoption of the term Carboniferous in 1822, Conybeare and Phillips referred to the French Terrain Crétacé as the Cretaceous System.

Clearly, surficial deposits and related unconsolidated material—variously relegated to the categories of classification proposed by Arduino, Lehmann, Werner, and others as "alluvium" or related formations—deserved a place in any formalized system of rock succession. In 1829 Jules Desnoyers of France, studying sediments in the Seine valley, proposed using the term Quaternary to encompass all of these various post-Tertiary formations. At nearly the same time, the important work of Lyell on the faunal succession of the Paris Basin permitted finer-scaled discrimination of this classic Tertiary sequence. In 1833 Lyell, using various biostratigraphic evidence, proposed several divisions of the Tertiary System that included the Eocene, Miocene, and Pliocene epochs. By 1839 he proposed using the term Pleistocene instead of dividing his Pliocene Epoch into older and newer phases. The temporal subdivision of the Tertiary was completed by two German scientists, Heinrich Ernst Beyrich and Wilhelm Philipp Schimper. Beyrich introduced the Oligocene in 1854 after having investigated outcrops in Belgium and Germany, while Schimper proposed adding the Paleocene in 1874 based on his studies of Paris Basin flora.

Werner's quadripartite division of rocks in southern Germany was applied well into the second decade of the 19th century. During this time, rock sequences from the lower part of his third temporal subdivision, the Flötzgebirge, were subsequently subdivided into three formations, each having fairly widespread exposure and distribution. Based on his earlier work, Friedrich August von Alberti identified in 1834 these three distinct lithostratigraphic units—the Bunter Sandstone, the Muschelkalk Limestone, and the Keuper Marls and Clays—as constituting the Trias or Triassic System.

Perhaps one of the most intriguing episodes in the development of the geologic time scale concerns the efforts of two British geologists and in large measure their attempts at unraveling the complex geologic history of Wales. Adam Sedgwick and Roderick Impey Murchison began working, in 1831, on the sequence of rocks lying beneath the Old Red Sandstone (which had been included in the basal sequence of the Carboniferous, as defined by Conybeare and Phillips, earlier in 1822). What started as an earnest collaborative attempt at deciphering the structurally and stratigraphically complicated rock succession in Wales ended in 1835 with a presentation outlining two distinct subdivisions of the pre-Carboniferous succession. Working up from the base of the post-Primary rock succession of poorly fossiliferous clastic rocks in northern Wales, Sedgwick identified a sequence of rock units defined primarily by their various lithologies. He designated this succession the Cambrian, after Cambria, the Roman name for Wales. Murchison worked downward in the considerably more fossiliferous pre-Old Red Sandstone rock sequence in southern Wales and was able to identify a succession of strata containing a well-preserved fossil fauna. These sequences defined from southern Wales were eventually brought into the context of Sedgwick's Cambrian. Murchison named his rock succession the Silurian, after the Roman name for an early Welsh tribe. In a relatively short time, Murchison's Silurian was expanding both laterally and temporally as more and more localities containing the characteristic Silurian fauna were recognized throughout Europe. The major problem created by this conceptual "expansion" of the Silurian was that it came to be recognized in northern Wales as coincident with much of the strata in the upper portion of Sedgwick's Cambrian. With Sedgwick's Cambrian based mainly on lithologic criteria, the presence of Silurian fauna created correlational difficulties. As it turned out, Sedgwick's Cambrian was of little value outside of its area of original definition. With it being superseded by the paleontologically based concept of the Silurian, some sort of compromise had to be worked out.

This compromise came about primarily as a result of the work of English geologist Charles Lapworth. In 1879 Lapworth proposed the designation Ordovician System for that sequence of rocks representing the upper part of Sedgwick's Cambrian succession and the lower (and generally overlapping) portion of Murchison's Silurian succession. The term Ordovician is derived from yet another Roman-named tribe of ancient Wales, the Ordovices. A large part of Lapworth's rationale for this division was based on the earlier work of the French-born geologist Joachim Barrande, who investigated the apparent Silurian fauna of central Bohemia. Barrande's 1851 treatise on this area of Czechoslovakia demonstrated a distinct succession from a "second" Silurian fauna to a "third" Silurian fauna. This divisible Silurian, as well as separate lines of evidence gathered by Lapworth in Scotland and Wales, finally enabled the individual character of the Cambrian, Ordovician, and Silurian systems to be resolved.

While involved in their work on Welsh stratigraphic successions, Sedgwick and Murchison had the opportunity to compare some rock outcroppings in Devonshire, in southwest England, with similar rocks in Wales. The Devon rocks were originally thought to belong to part of Sedgwick's Cambrian System, but they contained plant fossils very similar to basal Carboniferous (Old Red Sandstone) plant fossils found elsewhere. Eventually recognizing that these fossil-bearing sequences represented lateral equivalents in time and perhaps temporally unique strata as well, Sedgwick and Murchison in 1839 proposed the Devonian System.

During the early 1840s, Murchison traveled with the French paleontologist Edouard de Verneuil and the Latvian-born geologist Alexandr Keyserling to study the rock succession of the eastern Russian platform, the area of Russia west of the Ural Mountains. Near the town of Perm, Murchison and Verneuil identified fossiliferous strata containing both Carboniferous and a younger fauna at that time not recognized elsewhere in Europe or in the British Isles. Whereas the Carboniferous fossils were similar to those they had seen elsewhere (mainly from the Coal Measures), the stratigraphically higher fauna appeared somewhat transitional to the Triassic succession of Germany as then understood. Murchison coined the term Permian (after the town of Perm) to represent this intermediate succession.

With continued refinement of the definition of the Carboniferous in Europe, particularly in England, what at one time comprised the Old Red Sandstone, Lower Coal Measures (Mountain Limestone and Millstone Grit), and Upper Coal Measures now stood as just the Lower and Upper Coal Measures. It was beginning to be recognized that certain rock sequences in the Catskill Mountains of eastern New York State in North America resembled the Old Red Sandstone of western England. Furthermore, coal-bearing strata exposed in Pennsylvania greatly resembled the similar coal-bearing strata of the Upper Coal Measures. Lying beneath these coal-bearing rocks of Pennsylvania was a sequence of limestones that could be traced over thousands of square kilometres and that occurred in numerous outcrops along various tributary streams to the Ohio and Mississippi rivers in Indiana, Kentucky, Missouri, Illinois, and Iowa. This "subcarboniferous" strata, identified by the American geologist David Dale Owen in 1839, was subsequently termed Mississippian in 1870 as a result of work conducted by another American geologist, Alexander Winchell, in the upper Mississippi valley area. Eventually the overlying strata, the coal-bearing rocks originally described from Pennsylvania, were formalized as Pennsylvanian in 1891 by the paleontologist and stratigrapher Henry Shaler Williams.

The North American-defined Mississippian and Pennsylvanian systems were later correlated with presumed European and British successions. Although approximately similar in successional relationship, the Mississippian-Pennsylvanian boundary in North America is now considered slightly younger than the Lower–Upper Carboniferous boundary in Europe.

By the 1850s, with the development of the geologic time scale nearly complete, investigators were beginning to recognize that a number of major paleontologically defined boundaries were common and recurrent regardless of where a succession was studied. By this time rock successions were being defined according to fauna they contained, and the relative time scale, which was being erected, was based on the principle of faunal succession. Consequently, any major hiatus or change in faunal character was bound to be interpreted as important. In 1838 Sedgwick proposed that all pre-Old Red Sandstone sediments be included in the rock succession designated the Paleozoic Series (or Era) that contained generally primitive fossil fauna. John Phillips, another English geologist, went on to describe the Mesozoic Era to accommodate what then was the Cretaceous, Jurassic, Triassic, and partially Permian strata, and the Kainozoic (Cainozoic, or Cenozoic) Era to include Lyell's Eocene, Miocene, and Pliocene. This subdivision of the generally fossiliferous strata that lay superpositionally above the so-called Primary rocks of many of the early workers resulted in the recognition of three distinct eras. Subsequent subdivision of these eras into specific geologic periods finally provided the hierarchy for describing the relative dating of geologic events.

THE DEVELOPMENT OF RADIOACTIVE DATING METHODS AND THEIR APPLICATION

As has been seen, the geologic time scale is based on stratified rock assemblages that contain a fossil record. For the most part, these fossils allow various forms of information from the rock succession to be viewed in terms of their relative position in the sequence. Approximately the first 87 percent of Earth history occurred before the evolutionary development of shell-bearing organisms. The result of this mineralogic control on the preservability of organic remains in the rock record is that the geologic time scale—essentially a measure of biologic changes through time—takes in only the last 13 percent of Earth history. Although the span of time preceding the Cambrian period—the Precambrian—is nearly devoid of characteristic fossil remains and coincides with some of the primary rocks of certain early workers, it must, nevertheless, be evaluated in its temporal context.

Early Attempts at Calculating the Age of the Earth

Historically, the subdivision of Precambrian rock sequences (and, therefore, Precambrian time) had been accomplished on the basis of structural or lithologic grounds. With only minor indications of fossil occurrence (mainly in the form of algal stromatolites), no effective method of quantifying this loosely constructed chronology existed until the discovery of radioactivity enabled dating procedures to be applied directly to the rocks in question.

The quantification of geologic time remained an elusive matter for most human enquiry into the age of the Earth and its complex physical and biological history. Although Hindu teachings accept a very ancient origin for the Earth, medieval Western concepts of Earth history were based for the most part on a literal interpretation of Old Testament references. Biblical scholars of Renaissance Europe and later considered paternity as a viable method by which the age of the Earth since its creation could be determined. A number of attempts at using the "begat" method of determining the antiquity of an event—essentially counting backward in time through each documented human generation—led to the age of the Earth being calculated at several thousand years. One such attempt was made by Archbishop James Ussher of Ireland, who in 1650 determined that the Creation had occurred during the evening of Oct. 22, 4004 BCE. By his analysis of biblical genealogies, the Earth was not even 6,000 years old!

From the time of Hutton's refinement of uniformitarianism, the principle found wide application in various attempts to calculate the age of the Earth. As previously noted, fundamental to the principle was the premise that various Earth processes of the past operated in much the same way as those processes operate today. The corollary to this was that the rates of the various ancient processes could be considered the same as those of the present day. Therefore, it should be possible to calculate the age of the Earth on the basis of the accumulated record of some process that has occurred at this determinable rate since the Creation.

Many independent estimates of the age of the Earth have been proposed, each made using a different method of analysis. Some such estimates were based on assumptions concerning the rate at which dissolved salts or sediments are carried by rivers, supplied to the world's oceans, and allowed to accumulate over time. These chemical and physical arguments (or a combination of both) were all flawed to varying degrees because of an incomplete understanding of the processes involved. The notion that all of the salts dissolved in the oceans were the products of leaching from the land was first proposed by the English astronomer and mathematician Edmond Halley in 1691 and restated by the Irish geologist John Joly in 1899. It was assumed that the ocean was a closed system and that the salinity of the oceans was an ever-changing and ever-increasing condition. Based on these calculations, Joly proposed that the Earth had consolidated and that the oceans had been created between 80 and 90 million years ago. The subsequent recognition that the ocean is not closed and that a continual loss of salts occurs due to sedimentation in certain environments severely limited this novel approach.

Equally novel but similarly flawed was the assumption that, if a cumulative measure of all rock successions were compiled and known rates of sediment accumulation were considered, the amount of time elapsed could be calculated. While representing a reasonable approach to the problem, this procedure did not or could not take into account different accumulation rates associated with different environments or the fact that there are many breaks in the stratigraphic record. Even observations made on faunal succession proved that gaps in the record do occur. How long were these gaps? Do they represent periods of nondeposition or periods of deposition followed by periods of erosion? Clearly sufficient variability in a given stratigraphic record exists such that it may be virtually impossible to even come to an approximate estimate of the Earth's age based on this technique. Nevertheless, many attempts using this approach were made.

William Thomson (later Lord Kelvin) applied his thermodynamic principles to the problems of heat flow, and this had implications for predicting the age of a cooling

Sun and of a cooling Earth. From an initial estimate of 100 million years for the development of a solid crust around a molten core proposed in 1862, Thomson subsequently revised his estimate of the age of the Earth downward. Using the same criteria, he concluded in 1899 that the Earth was between 20 and 40 million years old.

British physicist and 1st Baron, Ernest Rutherford. Topical Press Agency/Hulton Archive/Getty Images

Thomson's calculation was based on the assumption that the substance of the Earth is inert and thus incapable of producing new heat. His estimate came into question after the discovery of naturally occurring radioactivity by the French physicist Henri Becquerel in 1896 and the subsequent recognition by his colleagues, Marie and Pierre Curie, that compounds of radium (which occur in uranium minerals) produce heat. As a result of this and other findings, notably that of Ernest Rutherford, it became apparent that naturally occurring radioactive elements in minerals common in the Earth's crust are sufficient to account for all observed heat flow. Within a short time, another leading British physicist, John William Strutt, concluded that the production of heat in the Earth's interior was a dynamic process—one in which heat was continuously provided by such materials as uranium. The Earth was, in effect, not cooling.

An Absolute Age Framework for the Stratigraphic Time Scale

In his book *Radio-activity* (1904), Rutherford explained that radioactivity results from the spontaneous disintegration of an unstable element into a lighter element, which may decay further until a stable element is finally created. This process of radioactive decay involves the emission of positively charged particles (later to be recognized as helium nuclei) and negatively charged ones (electrons) and in most cases gamma rays (a form of electromagnetic radiation) as well. This interpretation, the so-called disintegration theory, came to provide the basis for the numerical quantification of geologic time.

In 1905 Strutt succeeded in analyzing the helium content of a radium-containing rock and determined its age to be 2 billion years. This was the first successful application of a radiometric technique to the study of Earth materials, and it set the stage for a more complete analysis of geologic time. Although faced with problems of helium loss and therefore not quite accurate results, a major scientific breakthrough had been accomplished. Also in 1905 the American chemist Bertram B. Boltwood—working with the more stable uranium-lead system—calculated the numerical ages of 43 minerals. His results, with a range of 400 million to 2.2 billion years, were an order of magnitude greater than those of the other "quantitative" techniques of the day that made use of heat flow or sedimentation rates to estimate time.

Acceptance of these new ages was slow in coming. Perhaps much to their relief, paleontologists now had sufficient time in which to accommodate faunal change.

Researchers in other fields, however, were still conservatively sticking with ages on the order of several hundred million, but were revising their assumed sedimentation rates downward in order to make room for expanded time concepts.

In a brilliant contribution to resolving the controversy over the age of the Earth, Arthur Holmes, a student of Strutt, compared the relative (paleontologically determined) stratigraphic ages of certain specimens with their numerical ages as determined in the laboratory. This 1911 analysis provided for the first time the numerical ages for rocks from several Paleozoic geologic periods as well as from the Precambrian. Carboniferous-aged material was determined to be 340 million years, Devonian-aged material 370 million years, Ordovician (or Silurian) material 430 million years, and Precambrian specimens from 1.025 to 1.64 billion years. As a result of this work, the relative geologic time scale, which had taken nearly 200 years to evolve, could be numerically quantified. No longer did it have merely superpositional significance, it now had absolute temporal significance as well.

NONRADIOMETRIC DATING

In addition to radioactive decay, many other processes have been investigated for their potential usefulness in absolute dating. Unfortunately, they all occur at rates that lack the universal consistency of radioactive decay. Sometimes human observation can be maintained long enough to measure present rates of change, but it is not at all certain on a priori grounds whether such rates are representative of the past. This is where radioactive methods frequently supply information that may serve to calibrate nonradioactive processes so that they become useful chronometers. Nonradioactive absolute chronometers may conveniently be classified in terms of the broad areas in which changes occur—namely, geologic and biological processes, which will be treated here.

Geologic Processes as Absolute Chronometers

Rocks and minerals can be used in absolute dating. Wind, water, and other forces wear them down, and the particles that are removed become sediment. Over time, this sediment can accumulate in lake and ocean basins to become sedimentary rock, which can provide clues of how much time has passed. In addition, polar reversals occurring in Earth's magnetic field can leave their signatures in certain types of rocks.

Weathering Processes

During the first third of the 20th century, several presently obsolete weathering chronometers were explored. Most famous was the attempt to estimate the duration of Pleistocene interglacial intervals through depths of soil development. In the American Midwest, thicknesses of gumbotil and carbonate-leached zones were measured in the glacial deposits (tills) laid down during each of the four glacial stages. Based on a direct proportion between thickness and time, the three interglacial intervals were determined to be longer than postglacial time by factors of 3, 6, and 8. To convert these relative factors into absolute ages required an estimate in years of the length of postglacial time. When certain evidence suggested 25,000 years to be an appropriate figure, factors became years—namely, 75,000, 150,000, and 200,000 years. And, if glacial time and nonglacial time are assumed approximately equal, the Pleistocene Epoch lasted about 1 million years.

These glacial deposits (tills) are located in Glacier National Park, Montana. U.S. Geological Survey

Only one weathering chronometer is employed widely at the present time. Its record of time is the thin hydration layer at the surface of obsidian artifacts. Although no hydration layer appears on artifacts of the more common flint and chalcedony, obsidian is sufficiently widespread that the method has broad application.

In a specific environment the process of obsidian hydration is theoretically described by the equation $D = Kt^{1/2}$, in which D is thickness of the hydration rim, K is a constant characteristic of the environment, and t is the time since the surface examined was freshly exposed. This relationship is confirmed both by laboratory experiments at 100° C (212° F) and by rim measurements on obsidian artifacts found in carbon-14 dated sequences. Practical experience indicates that the constant K is almost totally dependent on temperature and that humidity is apparently of no significance. Whether in a dry Egyptian tomb or buried in wet tropical soil, a piece of obsidian seemingly has a surface that is saturated with a molecular film of water. Consequently, the key to absolute dating of obsidian is to evaluate *K* for different temperatures. Ages follow from the above equation provided there is accurate knowledge of a sample's temperature history. Even without such knowledge, hydration rims are useful for relative dating within a region of uniform climate.

Like most absolute chronometers, obsidian dating has its problems and limitations. Specimens that have been exposed to fire or to severe abrasion must be avoided. Furthermore, artifacts reused repeatedly do not give ages corresponding to the culture layer in which they were found but instead to an earlier time, when they were fashioned. Finally, there is the problem that layers may flake off beyond 40 micrometres (0.004 centimetre, or 0.002 inch) of thickness—that is, more than 50,000 years in age. Measuring several slices from the same specimen is wise in this regard, and such a procedure is recommended regardless of age.

Accumulational Processes

Sediment in former or present water bodies, salt dissolved in the ocean, and fluorine in bones are three kinds of natural accumulations and possible time indicators. To serve as geochronometers, the records must be complete and the accumulation rates known.

The fossiliferous part of the geologic column includes perhaps 122,000 metres (about 400,000 feet) of sedimentary rock if maximum thicknesses are selected from throughout the world. During the late 1800s, attempts were made to estimate the time over which it formed by assuming an average rate of sedimentation. Because there was great diversity among the rates assumed, the range of estimates was also large—from a high of 2.4 billion years to a low of 3 million years. In spite of this tremendous spread, most geologists felt that time in the hundreds of millions of years was necessary to explain the sedimentary record.

If the geologic column were made up entirely of annual layers, its duration would be easy to determine. Limited sedimentary deposits did accumulate in this way, and they are said to be varved. One year's worth of sediment is called a varve, and, in general, it includes two laminae per year.

Varves arise in response to seasonal changes. New Mexico's Castile Formation, for example, consists of alternating layers of gypsum and calcite that may reflect an annual temperature cycle in the hypersaline water from which the minerals precipitated. In moist, temperate climates, lake sediments collecting in the summer are richer in organic matter than those that settle during winter. This feature is beautifully seen in the seasonal progression of plant microfossils found in shales at Oensingen, Switz. In the thick oil shales of Wyoming and Colorado in the United States, the flora is not so well defined, but layers alternating in organic richness seem to communicate the same seasonal cycle. These so-called Green River Shales also contain abundant freshwater-fish fossils that confirm deposition in a lake. At their thickest, they span 792 vertical metres (2,600 feet). Because the average thickness of a varve is about 0.015 centimetre (0.006 inch), the lake is thought to have existed for more than 5 million years.

Each of the examples cited above is of a floating chronology—that is, a decipherable record of time that was terminated long ago. In Sweden, by contrast, it has been possible to tie a glacial varve chronology to present time, and so create a truly absolute dating technique. Where comparisons with radiocarbon dating are possible, there is general agreement.

As early as 1844, an English chemist named Middleton claimed that fossil bones contain fluorine in proportion to their antiquity. This idea is sound in principle, provided that all the other natural variables remain constant. Soil permeability, rainfall, temperature, and the concentration of fluorine in groundwater all vary with time and location, however. Fluorine dating is therefore not the simple procedure that Middleton envisioned.

Still, the idea that hydroxyapatite in buried bone undergoes gradual change to fluorapatite is a correct one. In a restricted locality where there is uniformity of climate and soil, the extent of fluorine addition is at least a measure of relative age and has been so used with notable success in dating certain hominid remains. Both the Piltdown hoax, for example, and the intrusive burial of the Galley Hill skeleton were exposed in part by fluorine measurements. Supplementing them were analyses of uranium, which resembles fluorine in its increase with time, and nitrogen, which decreases as bone protein decays away.

Fluorine changes could conceivably be calibrated if bone samples were found in a radiometrically dated sequence. Conditions governing fluorine uptake, however, are so variable even over short distances that it is risky to use fluorine content as an absolute chronometer much beyond the calibration site itself. In short, fluorine dating is not now and probably never will be an absolute chronometer. Even when used in relative dating, many fluorine analyses on diverse samples are needed, and these must be supplemented by uranium and nitrogen measurements to establish confidence in the chronological conclusions.

Geomagnetic Variations

Based on three centuries of direct measurement, the Earth's magnetic field is known to be varying slowly in both its intensity and direction. In fact, change seems to have been the rule throughout all of the Earth's past. Magnetic minerals in rocks (and in articles of fired clay) provide the record of ancient change, for they took on the magnetic field existing at the time of their creation or emplacement.

Polar reversals were originally discovered in lava rocks and since have been noted in deep-sea cores. In both cases the time dimension is added through radiometric methods applied to the same materials that show the reversals. The potassium-argon chronometer is the commonest chronometer used. A magnetic-polarity (or paleomagnetic) time scale has been proposed along the line of the geologic time scale; time divisions are called intervals, or epochs.

Biological Processes as Absolute Chronometers

Living or once-living things can also be used in absolute dating. However, their application is somewhat limited. The most well-known examples of biological processes that act as chronometers are tree-ring growth and coral growth. Tree rings of some species may be used to date some events within the last 8,000 years or so, whereas corals may be used to gauge the length of a year in the distant past.

Tree-Ring Growth

In the early 1900s, an American astronomer named Andrew E. Douglass went looking for terrestrial records of past sunspot cycles and not only found what he sought but also discovered a useful dating method in the process. The focus of his attention was the growth rings in trees—living trees, dead trees, beams in ancient structures, and even large lumps of charcoal.

The key documents for tree-ring dating, or dendrochronology, are those trees that grow or grew where roots receive water in direct proportion to precipitation. Under such a situation, the annual tree rings vary in width as a direct reflection of the moisture supplied. What is important in tree-ring dating is the sequence in which rings vary. Suppose, for example, that a 100-year-old tree is cut down and its ring widths are measured. The results can be expressed graphically, and, if a similar graph were made from a small stump found near the 100-year-old tree, the two graphs could be compared until a match of the curves was obtained. The time when the small stump was made would thereby be determined from the position of its outer ring alongside the 100-year record.

Not every tree species nor even every specimen of a suitable species can be used. In the American Southwest, success has been achieved with yellow pine, Douglas fir, and even sagebrush. Unfortunately, the giant sequoia of California does not live in a sufficiently sensitive environment to provide a useful record. The even older bristlecone pine in California's White Mountains does have a climate-sensitive record, but its area of growth is so limited and so inaccessible that no bristlecone specimens have so far appeared in archaeological sites. This shortcoming notwithstanding, dead bristlecone pine trees are presently providing rings as old as 8,200 years for dating by carbon-14. The purpose is to check the carbon-14 method.

Coral Growth

Certain fossil corals have long been used to date rocks relatively, but only recently has it been shown that corals may also serve as absolute geochronometers. They may do so by preserving a record of how many days there were in a year at the time they were growing. The number of days per year has decreased through time because the rate of rotation of the Earth has decreased. Geophysical evidence suggests that days are currently lengthening at the rate of 20 seconds per million years. If this were typical of the slowdown during the past, a year consisted of 423 days about 600 million years ago.

It is thought that horn corals indicate the number of days per year by means of their exceedingly fine external ridges of calcium carbonate, each of which is believed to represent a day's growth. Several hundred of the fine ridges also seem to cluster as a unit that presumably corresponds to one year. In certain modern West Indian corals, the number of fine ridges in a presumed annual increment is approximately 360, suggesting that coral patterns are being properly interpreted.

Not many fossil corals are in a state of preservation that permits the counting of ridges, but those that are seem to lend themselves well to this procedure. Several Middle Devonian corals indicate between 385 and 410 ridges, with an average of about 400. It remains to be seen whether this method of dating, so elegant in concept and so simple in application, will blossom or wither away in the years to come.

GEOCHRONOLOGY CONCEPTS

A number of terms are commonly used in the field of geochronology. Some of the more important concepts, such as "fossil" and the "Law of Faunal Succession," are listed below.

Bioherm

A bioherm is an ancient organic reef of moundlike form built by a variety of marine invertebrates, including corals, echinoderms, gastropods, mollusks, and others. Fossil calcareous algae are prominent in some bioherms. A structure built by similar organisms that is bedded but not moundlike is called a biostrome. Bioherms and biostromes occur in sedimentary rock strata of all geological ages, providing definitive information on paleoenvironments in the vicinity of their occurrence.

Biozone

Any stratigraphic unit consisting of all the strata containing a particular fossil and, hence, deposited during its existence is considered a biozone. The extent of the unit in a particular place, on the local stratigraphic range of the fossil plant or animal involved, is called a teilzone. The geological time units corresponding to biozones and teilzones are biochrons and teilchrons, respectively. Biozone is also used synonymously with the terms zone and range zone in stratigraphy.

Cordilleran Geosyncline

This linear trough in the Earth's crust—in which rocks of Late Precambrian to Mesozoic age (roughly 600 million to 66 million years ago) were deposited—spans the western coast of North America, from southern Alaska through western Canada and the United States, probably to western Mexico. The eastern boundary of the geosyncline extends from southeastern Alaska along the eastern edge of the Northern Cordillera and Northern Rocky Mountains of Canada and Montana, along the eastern edge of the Great Basin of Utah and Nevada, and into southeastern California and Mexico.

The principal mountain-building phases of the geosyncline took place during Mesozoic time, but many earlier orogenic events have also been recorded. Deformation of the Cordilleran Geosyncline and the formation of the Cordilleran fold appear to be related to the development of oceanic trenches along the western margins of the North American continent, the underriding of the continental plate by oceanic crust, and the development of batholithic intrusions and extrusion of volcanic rocks associated with this movement.

The Law of Faunal Succession

The Law of Faunal Succession derives from the observation that assemblages of fossil plants and animals follow or succeed each other in time in a predictable manner. Sequences of successive strata and their corresponding enclosed faunas have been matched together to form a composite section detailing the history of the Earth, especially from the inception of the Cambrian Period, which began about 540 million years ago. Faunal succession occurs because evolution generally progresses from simple to complex in a nonrepetitive and orderly manner. Because members of faunas can be distinguished from one another through time and because of the wide geographic distribution of organisms on the Earth, strata from different geographic areas can be correlated with each other and dated. Faunal succession is the fundamental tool of stratigraphy and comprises the basis for the geologic time scale. Climate and conditions throughout the Earth's history can be studied using the successive groups of plants and animals because they reflect their environment.

Faunizone

A faunizone is a stratigraphic unit that is distinguished by the presence of a particular fauna of some time or environmental significance. It differs from a biozone because it is based on a fossil assemblage rather than a particular genus or species. The corresponding unit of geologic time is called a faunichron.

Fossil

A fossil is a remnant, impression, or trace of an animal or plant of a past geologic age that has been preserved in the Earth's crust. The complex of data recorded in fossils worldwide, known as the fossil record, is the primary source of information about the history of life on Earth.

Only a small fraction of ancient organisms are preserved as fossils, and usually only organisms that have a solid and resistant skeleton are readily preserved. Most major groups of invertebrate animals have a calcareous skeleton or shell (such as corals, mollusks, brachiopods, bryozoans). Other forms have shells of calcium phosphate (which also occurs in the bones of vertebrates), or silicon dioxide. A shell or bone that is buried quickly after

Fossilized footprint of an unidentified dinosaur. © Getty Images deposition may retain these organic tissues, though they become petrified (converted to a stony substance) over time. Unaltered hard parts, such as the shells of clams or brachiopods, are relatively common in sedimentary rocks, some of great age.

The hard parts of organisms that become buried in sediment may be subject to a variety of other changes during their conversion to solid rock, however. Solutions may fill the interstices, or pores, of the shell or bone with calcium carbonate or other mineral salts and thus fossilize the remains, in a process known as permineralization. In other cases there may be a total replacement of the original skeletal material by other mineral matter, a process known as mineralization, or replacement. In still other cases, circulating acid solutions may dissolve the original shell but leave a cavity corresponding to it, and circulating calcareous or siliceous solutions may then deposit a new matrix in the cavity, thus creating a new impression of the original shell.

Dinosaur fossils found in Alberta, Canada. AbleStock/Jupiterimages

By contrast, the soft parts of animals or plants are very rarely preserved. The embedding of insects in amber and the preservation of the carcasses of Pleistocene-era mammoths in ice are rare but striking examples of the fossil preservation of soft tissues. Traces of organisms may also occur as tracks or trails or even borings.

The great majority of fossils are preserved in a water environment because land remains are more easily destroyed. Anaerobic conditions at the bottom of the seas or other bodies of water are especially favourable for preserving fine details, since no bottom faunas, except for anaerobic bacteria, are present to destroy the remains. In general, for an organism to be preserved two conditions must be met: rapid burial to retard decomposition and to prevent the ravaging of scavengers; and possession of hard parts capable of being fossilized.

In some places, such as the Grand Canyon in northern Arizona, one can observe a great thickness of nearly horizontal strata representing the deposition of sediment on the seafloor over many hundreds of millions of years. It is often apparent that each layer in such a sequence contains fossils that are distinct from those of the layers that are above and below it. In such sequences of layers in different geographic locations, the same, or similar, fossil floras or faunas occur in the identical order. By comparing overlapping sequences, it is possible to build up a continuous record of faunas and floras that have progressively more in common with present-day life-forms as the top of the sequence is approached.

Fossils provide the geologist a quick and easy way of assigning a relative age to the strata in which they occur. The precision with which this may be done in any particular case depends on the nature and abundance of the fauna: some fossil groups were deposited during much longer time intervals than others. Fossils used to identify geologic relationships are known as index fossils.

Fossil organisms, furthermore, may provide useful information about the climate and environment of the site where they were deposited and preserved. Certain species of coral, for example, require warm, shallow water; certain plants require warm, swampy conditions such as are found today in the Florida Everglades. Thus, when rocks containing fossils of this kind are found in rocks of the present-day polar regions, there is a strong presumption that the crust on which they were deposited has shifted thousands of miles since that time.

Fossilized leaf. PhotoObjects.net/ Jupiterimages

Fossils are useful in the exploration for minerals and mineral fuels. For example, they serve to indicate the stratigraphic position of coal seams. In recent years, geologists have been able to study the subsurface stratigraphy of oil and natural gas deposits by analyzing microfossils obtained from core samples of deep borings.

Fossil collection performed by paleontologists, geologists, and other scientists typically involves a rigorous excavation and documentation process. Unearthing the specimen from the rock often involves a painstaking process in which the parts are labeled and catalogued by location within the rock. Those fossils slated for removal from the rock are slowly and carefully excavated using techniques designed to prevent and minimize damage to the specimen.

Many fossils, however, are collected by hobbyists and commercial entities. Often such specimens are not carefully documented or excavated, resulting in a loss of data from the site and risking potential damage to the specimen. For these reasons and the fact that it stimulates nonscientific collecting, the commercial exploitation of fossils is controversial among academic paleontologists.

Fossil Record

The fossil record is the history of life as documented by the remains or imprints of the organisms from earlier geological periods preserved in sedimentary rock. In a few cases, the original substance of the hard parts of the organism is preserved, but more often the original components have been replaced by minerals deposited from water seeping through the rock. Occasionally the original material is simply removed, while nothing is deposited in its place. In this case, all that remains is a mould of the shape of the plant or animal.

Study of the fossil record has provided important information for at least three different purposes. The progressive changes observed within an animal group are used to describe the evolution of that group. In general, but not always, successive generations tend to change morphologically in a particular direction (such as the progressive acquisition or loss of specific features), and these changes are often interpreted as better adaptation (through preferential selection of beneficial mutations) to a particular environment.

Index Fossil

An index fossil is any animal or plant preserved in the rock record of the Earth that is characteristic of a particular span of geologic time or environment. A useful index fossil must be distinctive or easily recognizable, be abundant, and have a wide geographic distribution and a short range through time. Index fossils are the basis for defining boundaries in the geologic time scale and for the

Any widely distributed fossils that are characteristic of a certain age, such as ammonites are in the Jurassic Era, can serve as index fossils. Shutterstock.com correlation of strata. In marine strata, index fossils that are commonly used include the single-celled Protista with hard body parts and larger forms such as ammonoids. In terrestrial sediments of the Cenozoic Era, which began about 65.5 million years ago, mammals are widely used to date deposits. All of these animal forms have hard body parts, such as shells, bones, and teeth, and evolved rapidly.

Marker Bed

A marker, or key, bed is a rock unit that is readily distinguishable by reason of physical characteristics and traceable over large horizontal distances. Stratigraphic examples include coal beds and beds of volcanic ash. The term marker bed is also applied to sedimentary strata that provide distinctive seismic reflections.

Paleogeology

Paleogeology is the geology of a region at any given time in the distant past. Paleogeologic reconstructions in map form show not only the ancient topography of a region but also the distribution of rocks beneath the surface and such structural features as faults and folds. Maps of this kind help investigators to better determine the instances of deformation events in a region, the stream-drainage patterns now buried under layers of sediment, and the extent of ancient oceans and seas. They also provide a useful tool for petroleum geologists, enabling them to identify geologic structures where oil or natural gas may be trapped.

Remanent Magnetism

Remanent magnetism, or paleomagnetism, is the permanent magnetism in rocks, resulting from the orientation of the Earth's magnetic field at the time of rock formation in a past geological age. It is the source of information for the paleomagnetic studies of polar wandering and continental drift. Remanent magnetism can derive from several natural processes, generally termed natural remanent magnetism, the most important being thermo-remanent magnetism. This arises when magnetic minerals forming in igneous rocks cool through the Curie point and when the magnetic domains within the individual minerals align themselves with the Earth's magnetic field, thus making a permanent record of its orientation.

A second mechanism operates when small grains of magnetic minerals settle into a sedimentary matrix, producing detrital remanent magnetism. It is hypothesized that the tiny grains orient themselves in the direction of the Earth's magnetic field during deposition and before the final consolidation of the rock. The magnetism thus introduced appears to persist through later alteration and compaction of the rock, although the details of these processes have not been fully studied.

Rocks may acquire remanent magnetism in at least two other ways: (1) rocks made up of nonmagnetic minerals may be chemically altered to yield magnetic minerals, and these newly formed minerals will acquire remanent magnetism in the presence of the Earth's magnetic field; and (2) igneous rocks already cooled may ultimately acquire remanent magnetism by a process called viscous magnetization. The difference between these several types of remanent magnetism can be determined, and the magnetic history of a particular rock can therefore be interpreted.

Tephrochronology

The method of age determination that makes use of layers of ash (tephra) is called tephrochronology. Tephra layers are excellent time-stratigraphic markers, but, to establish a chronology, it is necessary to identify and correlate as many tephra units as possible over the widest possible area. Because of the large number of violent volcanic explosions in Iceland, Sigurdur Thorarinsson—an Icelander who was the founder of the science of tephrochronology—was able to establish a detailed chronology of preoccupational and postoccupational geologic and archaeological events there. Tephrochronology enabled Thorarinsson to make a thorough study of the changes in climate in Iceland and the effect of agriculture on Icelandic ecology.

Japan is another fertile field for tephrochronological studies, and, not surprisingly, Japanese scientists have greatly contributed to the development of the science. In western North America, Pleistocene tephra (those from 11,700 to 2.6 million years old) have been correlated over a distance greater than 1,600 km (1,000 miles). Tephra may be traced from the land areas into the ocean basins and, thus, may provide one of the most effective techniques for correlating terrigenous with marine stratigraphic deposits.

Varved Deposit

A varved deposit is any form of repetitive sedimentary rock stratification, either bed or lamination, that was deposited within a one-year time period. This annual deposit may comprise paired contrasting laminations of alternately finer and coarser silt or clay, reflecting seasonal sedimentation (summer and winter) within the year. Varved deposits are to be distinguished from rhythmites, the latter also being made up of paired laminations or beds but with an annual cyclicity that cannot be proved.

Varved deposits are usually associated with fine-grained sediments, the muds or mudrocks, which include both silt- and clay-grade materials. Laminations in many mudrocks are both thin and laterally persistent over large areas. They may exhibit the right order of thickness, as shown by the rates of sedimentation estimated for times past or observed at present, and have a structure similar to laminations currently being formed. Using this premise, one may deduce that many mudrock laminations are of an annual nature and that the varved deposits depend upon the yearly climatic cycle. This cycle affects temperature, salinity, and silt content of waters as well as the seasonal production of plankton.

Varved deposits are most commonly associated with sedimentation in lakes, particularly those that are located in glacial or proglacial environments. During the summer months, sediment is transported into the lake from the surrounding drainage basin as a result of ice melting and outwash. The central part of the lake receives relatively coarse sediment—silt to very fine-grade sand detritus—distributed by currents. The sediment settles to the bottom, with the coarser particles settling faster. In the wintertime the lake may not receive any new sediment input, probably because the lake is ice covered. Thus, the finest sediments, the clays, flocculate in the water column and settle out of suspension in the lake. The end product is a coarse- and fine-grained sediment couplet, the sediment being light (summer) and dark coloured (winter), respectively. This couplet is the hallmark of varved deposits. The annual cyclicity of the varved deposits in modern lakes can be proven as seasonal by using pollen analysis or by undertaking carbon-14 dating of the succession. Some varved sediments in the glacigenic environment can display an exponential decrease in thickness of the couplet away from the ice front. This may be a result, in part, of the couplet being deposited by density current and autosuspension mechanisms operating within the water body.

Varved deposits in recent and ancient sedimentary sequences, where they are often termed varvite, frequently display disruption of the fine lamination and couplets by outsize clasts. These clasts are called dropstones and were introduced vertically through the water column into the lake area, where only fine-grained sediments normally accumulate, by ice rafting and melting. This phenomenon of disrupted varvites constitutes the strongest evidence of past glacial activity in a region.

Varved sediments also can be found in nonglacial lakes and marine settings and as a result of aeolian processes. Varved deposits are commonly associated with evaporite sequences where lake sediments display detrital and chemical cycles. These thicker couplets are the product of a cyclic variation in rainfall imposed upon the deposits of a continuously subsiding lake basin.

CHAPTER 2

DATING

In geology, dating is the process that determines a chronology or calendar of events in the history of the Earth, using to a large degree the evidence of organic evolution in the sedimentary rocks accumulated through geologic time in marine and continental environments. To date past events, processes, formations, and fossil organisms, geologists employ a variety of techniques. These include some that establish a relative chronology in which occurrences can be placed in the correct sequence relative to one another or to some known succession of events. Radiometric dating and certain other approaches are used to provide absolute chronologies in terms of years before the present. The two approaches are often complementary, as when a sequence of occurrences in one context can be correlated with an absolute chronlogy elsewhere.

THE DISTINCTIONS BETWEEN RELATIVE-AGE AND ABSOLUTE-AGE MEASUREMENTS

Local relationships on a single outcrop or archaeological site can often be interpreted to deduce the sequence in which the materials were assembled. This then can be used to deduce the sequence of events and processes that took place or the history of that brief period of time as recorded in the rocks or soil. For example, the presence of recycled bricks at an archaeological site indicates the sequence in which the structures were built. Similarly, in geology, if distinctive granitic pebbles can be found in the sediment beside a similar granitic body, it can be inferred that the granite, after cooling, had been uplifted and eroded and therefore was not injected into the adjacent rock sequence. Although with clever detective work, many complex time sequences or relative ages can be deduced. The ability to show that objects at two separated sites were formed at the same time requires additional information. A coin, vessel, or other common artifact could link two archaeological sites, but the possibility of recycling would have to be considered. It should be emphasized that linking sites together is essential if the nature of an ancient society is to be understood, as the information at a single location may be relatively insignificant by itself. Similarly, in geologic studies, vast quantities of information from widely spaced outcrops have to be integrated. Some method of correlating rock units must be found. In the ideal case, the geologist will discover a single rock unit with a unique collection of easily observed attributes called a marker horizon that can be found at widely spaced localities. Any feature, including colour variations, textures, fossil content, mineralogy, or any unusual combinations of these can be used. It is only by correlations that the conditions on different parts of the Earth at any particular stage in its history can be deduced. In addition, because sediment deposition is not continuous and much rock material has been removed by erosion, the fossil record from many localities has to be integrated before a complete picture of the evolution of life on Earth can be assembled. Using this established record, geologists have been able to piece together events over the past 600 million years, or about one-eighth of Earth history, during which time useful fossils have been abundant. The need to correlate over the rest of geologic time, to correlate nonfossiliferous units, and to calibrate the fossil time scale has led to the development of a specialized field that makes use of natural radioactive isotopes in order to calculate absolute ages.

The precise measure of geologic time has proven to be the essential tool for correlating the global tectonic processes that have taken place in the past. Precise isotopic ages are called absolute ages, since they date the timing of events not relative to each other but as the time elapsed between a rock-forming event and the present. Absolute dating by means of uranium and lead isotopes has been improved to the point that for rocks 3 billion years old geologically meaningful errors of ± 1 or 2 million years can be obtained. The same margin of error applies for younger fossiliferous rocks, making absolute dating comparable in precision to that attained using fossils. To achieve this precision, geochronologists have had to develop the ability to isolate certain high-quality minerals that can be shown to have remained closed to migration of the radioactive parent atoms they contain and the daughter atoms formed by radioactive decay over billions of years of geologic time. In addition, they have had to develop special techniques with which to dissolve these highly refractory minerals without contaminating the small amount (about one-billionth of a gram) of contained lead and uranium on which the age must be calculated. Since parent uranium atoms change into daughter atoms with time at a known rate, their relative abundance leads directly to the absolute age of the host mineral. Just as the use of the fossil record has allowed a precise definition of geologic processes in approximately the past 600 million years, absolute ages allow correlations back to the Earth's oldest known rocks formed almost 4 billion years ago. In fact, even in younger rocks, absolute dating is the only way that the fossil record can be calibrated. Without absolute ages, investigators could only determine which fossil organisms lived at the same time and the relative order of their appearance in the correlated sedimentary rock record.

Unlike ages derived from fossils, which occur only in sedimentary rocks, absolute ages are obtained from minerals that grow as liquid rock bodies cool at or below the surface. When rocks are subjected to high temperatures and pressures in mountain roots formed where continents collide, certain datable minerals grow and even regrow to record the timing of such geologic events. When these regions are later exposed in uptilted portions of ancient continents, a history of terrestrial rock-forming events can be deduced. Episodes of global volcanic activity, rifting of continents, folding, and metamorphism are defined by absolute ages. The results suggest that the present-day global tectonic scheme was operative in the distant past as well.

THE GLOBAL TECTONIC ROCK CYCLE

Bringing together virtually all geologic aspects of the Earth's outer rock shell (the lithosphere) into a unifying theory called plate tectonics has had a profound impact on the scientific understanding of our dynamic planet. Continents move, carried on huge slabs, or plates, of dense rock about 100 kilometres (about 62 miles) thick over a low-friction, partially melted zone (the asthenosphere) below. In the oceans, new seafloor, created at the globe-circling oceanic ridges, moves away, cools, and sinks back into the mantle in what are known as subduction zones (long, narrow belts at which one plate descends beneath another). Where this occurs at the edge of a continent, as along the west coast of North and South America, large mountain chains develop with abundant volcanoes and their subvolcanic equivalents. These units, called igneous rock, or magma in their molten form, constitute major crustal additions. By contrast, crustal destruction occurs at the margins of two colliding continents, as, for example, where the subcontinent of India is moving north over Asia. Great uplift, accompanied by rapid erosion, is taking place and large sediment fans are being deposited in the Indian Ocean to the south. With time, water-soluble "cement" will cause the sandy units to become sandstone. Rocks of this kind in the ancient record may very well have resulted from rapid uplift and continent collision.

When continental plates collide, the edge of one plate is thrust onto that of the other. The rocks in the lower slab undergo changes in their mineral content in response to heat and pressure and will probably become exposed at the surface again some time later. Rocks converted to new mineral assemblages because of changing temperatures and pressures are called metamorphic. Virtually any rock now seen forming at the surface can be found in exposed deep crustal sections in a form that reveals through its mineral content the temperature and pressure of burial. Such regions of the crust may even undergo melting and subsequent extrusion of melt magma, which may appear at the surface as volcanic rocks or may solidify as it rises to form granites at high crustal levels. Magmas produced in this way are regarded as recycled crust, whereas others extracted by partial melting of the mantle below are considered primary.

Even the oceans and atmosphere are involved in this great cycle because minerals formed at high temperatures are unstable at surface conditions and eventually break down or weather, in many cases taking up water and carbon dioxide to make new minerals. If such minerals were deposited on a downgoing (subducted) oceanic slab, they would eventually be heated and changed back into high-temperature minerals, with their volatile components being released. These components would then rise and be fixed in the upper crust or perhaps reemerge at the surface. Such hot circulating fluids can dissolve metals and eventually deposit them as economic mineral deposits on their way to the surface.

Geochronological studies have provided documentary evidence that these rock-forming and rock-re-forming processes were active in the past. Seafloor spreading has been traced, by dating minerals found in a unique grouping of rock units thought to have been formed at the oceanic ridges, to 500 million years ago, with rare occurrences as early as 2 billion years ago. Volcanic units resembling those formed over oceanic subduction zones can be dated worldwide to show that the Earth's most prolific volcanic event occurred about 2.7 billion years ago. Other ancient volcanic units document various cycles of mountain building. The source of ancient sediment packages like those presently forming off India can be identified by dating single detrital grains of zircon found in sandstone. Magmas produced by the melting of older crust can be identified because their zircons commonly contain inherited older cores. Episodes of continental collision can be dated by isolating new zircons formed as the buried rocks underwent local melting. Periods of deformation associated with major collisions cannot be directly dated if no new minerals have formed. The time of deformation can be bracketed, however, if datable units, which both predate and postdate it, can be identified. The timing of cycles involving the expulsion of fluids from deep within the crust can be ascertained by dating new minerals formed at high pressures in exposed deep crustal sections. In some cases, it is possible to prove that gold deposits may have come from specific fluids if the deposition time of the deposits can be determined and the time of fluid expulsion is known.

Where the crust is under tension, as in Iceland, great fissures develop. These fissures serve as conduits that allow black lava, called basalt, to reach the surface. The portion that remains in a fissure below the surface usually forms a vertical black tubular body known as a dike (or dyke). Precise dating of such dikes can reveal times of crustal rifting in the past. Dikes and lava, now exposed on either side of Baffin Bay, have been dated to determine the time when Greenland separated from North America—namely, about 60 million years ago.

Combining knowledge of the Earth processes observed today with absolute ages of ancient geologic analogues seems to indicate that the oceans and atmosphere were present by at least 3.5 billion years ago and that they were probably released by early heating of the planet. The continents were produced over time; the earliest portions were formed nearly 4 billion years ago, and the process still continues today. Absolute dating allows rock units formed at the same time to be identified and reassembled into ancient mountain belts, which in many cases have been disassociated by subsequent tectonic processes. The most obvious of these is the Appalachian chain that occupies the east coast of North America and extends to parts of Newfoundland as well as parts of Ireland, England, and Norway. Relic oceanic crust, formed between 480 and 500 million years ago, was identified on both sides of the Atlantic in this chain, as were numerous correlative volcanic and sedimentary units. Evidence based on geologic description, fossil content, and absolute and relative ages leave no doubt that these rocks were all part of a single mountain belt before the Atlantic Ocean opened in stages from about 200 million years ago.

THE DETERMINATION OF SEQUENCE

Relative geologic ages can be deduced in rock sequences consisting of sedimentary, metamorphic, or igneous rock units. In fact, they constitute an essential part in any precise isotopic, or absolute, dating program. Such is the case because most rocks simply cannot be isotopically dated. Therefore, a geologist must first determine relative ages and then locate the most favourable units for absolute dating. It is also important to note that relative ages are inherently more precise, since two or more units deposited minutes or years apart would have identical absolute ages but precisely defined relative ages. While absolute ages require expensive, complex analytical equipment, relative ages can be deduced from simple visual observations.

Most methods for determining relative geologic ages are well illustrated in sedimentary rocks. These rocks cover roughly 75 percent of the surface area of the continents, and unconsolidated sediments blanket most of the ocean floor. They provide evidence of former surface conditions and the life-forms that existed under those conditions. The sequence of a layered sedimentary series is easily defined because deposition always proceeds from the bottom to the top. This principle would seem self-evident, but its first enunciation more than 300 years ago by Nicolaus Steno represented an enormous advance in understanding. Known as the principle of superposition, it holds that in a series of sedimentary layers or superposed lava flows the oldest layer is at the bottom, and layers from there upward become progressively younger. On occasion, however, deformation may have caused the rocks of the crust to tilt, perhaps to the point of overturning them. Moreover, if erosion has blurred the record by removing substantial portions of the deformed sedimentary rock, it may not be at all clear which edge of a given layer is the original top and which is the original bottom.

Identifying top and bottom is clearly important in sequence determination, so important in fact that a considerable literature has been devoted to this question alone. Many of the criteria of top and bottom determination are based on asymmetry in depositional features. Oscillation ripple marks, for example, are produced in sediments by water sloshing back and forth. When such marks are preserved in sedimentary rocks, they define the original top and bottom by their asymmetric pattern. Certain fossils also accumulate in a distinctive pattern or position that serves to define the top side.

This sandstone bed in the Cretaceous Dakota formation in Colorado shows oscillation ripple marks. Dr. Marli Miller/Visuals Unlimited/Getty Images

In wind-blown or water-lain sandstone, a form of erosion during deposition of shifting sand removes the tops of mounds to produce what are called cross-beds. The truncated layers provide an easily determined depositional top direction. The direction of the opening of mud cracks or rain prints can indicate the uppermost surface of mudstones formed in tidal areas. When a section of rock is uplifted and eroded, as during mountain-building episodes, great volumes of rock are removed, exposing a variety of differently folded and deformed rock units. The new erosion surface must postdate all units, dikes, veins, and deformation features that it crosses. Even the shapes formed on the erosional or depositional surfaces of the ancient seafloor can be used to tell which way was up. A fragment broken from one bed can only be located in a younger unit, and a pebble or animal track can only deform a preexisting unit—that is, the one below. In fact, the number of ways in which one can determine the tops of well-preserved sediments is limited only by the imagination, and visual criteria can be deduced by amateurs and professionals alike.

One factor that can upset the law of superposition in major sediment packages in mountain belts is the presence of thrust faults. Such faults, which are common in compression zones along continental edges, may follow bedding planes and then cross the strata at a steep angle, placing older units on top of younger ones. In certain places, the fault planes are only a few centimetres thick and are almost impossible to detect.

Relative ages also can be deduced in metamorphic rocks as new minerals form at the expense of older ones in response to changing temperatures and pressures. In deep mountain roots, rocks can even flow like toothpaste in their red-hot state. Local melting may occur, and certain minerals suitable for precise isotopic dating may form both in the melt and in the host rock. In the latter case, refractory grains in particular may record the original age of the rock in their cores and the time of melting in their newly grown tips. Analytical methods are now available to date both growth stages, even though each part may weigh only a few millionths of a gram. Rocks that flow in a plastic state record their deformation in the alignment of their constituent minerals. Such rocks then predate the deformation. If other rocks that are clearly not deformed can be found at the same site, the time of deformation can be inferred to lie between the absolute isotopic ages of the two units.

Igneous rocks provide perhaps the most striking examples of relative ages. Magma, formed by melting deep within the Earth, cuts across and hence postdates all units as it rises through the crust, perhaps even to emerge at the surface as lava. Black lava, or basalt, the most common volcanic rock on Earth, provides a simple means for determining the depositional tops of rock sequences as well as proof of the antiquity of the oceans. Pillow shapes are formed as basaltic lava is extruded (erupted) under water. These are convex upward with a lower tip that projects down between two convex tops below. The shapes of pillows in ancient basalts provide both a direct indication of depositional top and proof of underwater eruption. They are widespread in rocks as old as 3.5 billion years, implying that the oceans were already present.

Basaltic lava rocks that are common where ancient continents have been rifted apart are fed from below by near vertical fractures penetrating the crust. Material that solidifies in such cracks remains behind as dikes. Here, the dikes must be younger than all other units. A more interesting case develops when a cooled older crust is fractured, invaded by a swarm of dikes, and subsequently subjected to a major episode of heating with deformation and intrusion of new magma. In this instance, even though the resulting outcrop pattern is extremely complex, all of the predike units can be distinguished by the relic dikes present. The dikes also record in their newly formed minerals components that can be analyzed to give both the absolute age and the temperature and pressure of the second event. Because dike swarms are commonly widespread, the conditions determined can often be extrapolated over a broad region. Dikes do not always continue upward in a simple fashion. In some cases, they spread between the layers of near-horizontal sedimentary or volcanic units to form bodies called sills. In this situation, fragments of the host rock must be found within the intrusive body to establish its relatively younger age.

Once most or all of the relative ages of various strata have been determined in a region, it may be possible to deduce that certain units have been offset by movement along fractures or faults while others have not. Dikes that cross fault boundaries may even be found. Application of the simple principle of crosscutting relationships can allow the relative ages of all units to be deduced.

The principles for relative age dating described above require no special equipment and can be applied by anyone on a local or regional scale. They are based on visual observations and simple logical deductions and rely on a correlation and integration of data that occurs in fragmentary form at many outcrop locations.

CORRELATION

Correlation is, as mentioned earlier, the technique of piecing together the informational content of separated outcrops. When information derived from two outcrops is integrated, the time interval they represent is probably greater than that of each alone. Presumably if all the world's outcrops were integrated, sediments representing all of geologic time would be available for examination. This optimistic hope, however, must be tempered by the realization that much of the Precambrian record—older than 542 million years—is missing.

Principles and Techniques

Correlating two separated outcrops means establishing that they share certain characteristics indicative of contemporary formation. The most useful indication of time equivalence is similar fossil content, provided of course that such remains are present. The basis for assuming that like fossils indicate contemporary formation is faunal succession. However, as previously noted, times of volcanism and metamorphism, which are both critical parts of global processes, cannot be correlated by fossil content. Furthermore, useful fossils are either rare or totally absent in rocks from Precambrian time, which constitutes more than 87 percent of Earth history. Precambrian rocks must therefore be correlated by means of precise isotopic dating.

Unlike the principles of superposition and crosscutting, faunal succession is a secondary principle. That is to say, it depends on other sequence-determining principles for establishing its validity. Suppose there exist a number of fossil-bearing outcrops each composed of sedimentary layers that can be arranged in relative order, primarily based on superposition. Suppose, too, that all the layers contain a good representation of the animal life existing at the time of deposition. From an examination of such outcrops with special focus on the sequence of animal forms comes the empirical generalization that the faunas of the past have followed a specific order of succession, and so the relative age of a fossiliferous rock is indicated by the types of fossils it contains.

English geologist William Smith. Photos.com

English engineer and geologist William Smith first noticed around 1800 that the different rock layers he encountered in his work were characterized by different fossil assemblages. Using fossils simply for identification purposes, Smith constructed a map of the various surface rocks outcropping throughout England, Wales, and southern Scotland. Smith's geologic map was extremely crude, but in its effect on Earth study, it was a milestone.

Following Smith's pioneering work, generations of geologists have confirmed that similar and even more extensive fossil sequences exist elsewhere. To this day, fossils are useful as correlation tools to geologists specializing in stratigraphy. In dating the past, the primary value of fossils lies within the principle of faunal succession: each interval of geologic history had a unique fauna that associates a given fossiliferous rock with that particular interval.

The basic conceptual tool for correlation by fossils is the index, or guide, fossil. Ideally, an index fossil should be such as to guarantee that its presence in two separated rocks indicates their synchroneity. This requires that the lifespan of the fossil species be but a moment of time relative to the immensity of geologic history. In other words, the fossil species must have had a short temporal range. On the practical side, an index fossil should be distinctive in appearance so as to prevent misidentification, and it should be cosmopolitan both as to geography and as to rock type. In addition, its fossilized population should be sufficiently abundant for discovery to be highly probable. Such an array of attributes represents an ideal, and much stratigraphic geology is rendered difficult because of departure of the natural fossil assemblage from this ideal. Nevertheless, there is no greater testimony to the validity of fossil-based stratigraphic geology than the absolute dates made possible through radioactive measurements. Almost without exception, the relative order of strata defined by fossils has been confirmed by radiometric ages.

Correlation based on the physical features of the rock record also has been used with some success, but it is restricted to small areas that generally extend no more than several hundred kilometres. The first step is determining whether similar beds in separated outcrops can actually be traced laterally until they are seen to be part of the same original layer. Failing that, the repetition of a certain layered sequence (such as a black shale sandwiched between a red sandstone and a white limestone) lends confidence to physical correlation. Finally, the measurement of a host of rock properties may well be the ultimate key to correlation of separated outcrops. The more ways in which two rocks are physically alike, the more likely it is that the two formed at the same time.

Only a partial listing of physical characteristics is necessary to indicate the breadth of approach in this area. Such features as colour, ripple marks, mud cracks, raindrop imprints, and slump structures are directly observable in the field. Properties derived from laboratory study include (1) size, shape, surface appearance, and degree of sorting of mineral grains, (2) specific mineral types present and their abundances, (3) elemental composition of the rock as a whole and of individual mineral components, (4) type and abundance of cementing agent, and (5) density, radioactivity, and electrical-magnetic-optical properties of the rock as a whole.

With the development of miniaturized analytical equipment, evaluation of rock properties down a small drill hole has become possible. The technique, called well logging, involves lowering a small instrument down a drill hole on the end of a wire and making measurements continuously as the wire is played out in measured lengths. By this technique it is possible to detect depth variations in electrical resistivity, self-potential, and gamma-ray emission rate and to interpret such data in terms of continuity of the layering between holes. Subsurface structures can thus be defined by the correlation of such properties.

Field geologists always prize a layer that is so distinctive in appearance that a series of tests need not be made to establish its identity. Such a layer is called a key bed. In a large number of cases, key beds originated as volcanic ash. Besides being distinctive, a volcanic-ash layer has four other advantages for purposes of correlation: it was laid down in an instant of geologic time; it settles out over tremendous areas; it permits physical correlation between contrasting sedimentary environments; and unaltered mineral crystals that permit radiometric measurements of absolute age often are present.

Correlation may be difficult or erroneous if several different ash eruptions occurred, and a layer deposited in one is correlated with that from another. Even then, the correlation may be justified if the two ash deposits represent the same volcanic episode. Much work has been undertaken to characterize ash layers both physically and chemically and so avoid incorrect correlations. Moreover, single or multigrain zircon fractions from the volcanic source are now being analyzed to provide precise absolute ages for the volcanic ash and the fossils in the adjacent units.

Geologic Column and its Associated Time Scale

The end product of correlation is a mental abstraction called the geologic column. It is the result of integrating all the world's individual rock sequences into a single sequence. In order to communicate the fine structure of this so-called column, it has been subdivided into smaller units. Lines are drawn on the basis of either significant changes in fossil forms or discontinuities in the rock record (unconformities, or large gaps in the sedimentary sequence). The basic subdivisions of rock are called systems, and the corresponding time intervals are termed periods. In the upper part of the geologic column, where fossils abound, these rock systems and geologic periods are the basic units of rock and time. Lumping of periods results in eras, and splitting gives rise to epochs. In both cases, a threefold division into early–middle–late is often used, although those specific words are not always applied. Similarly, many periods are split into three epochs. Names assigned to individual epochs follow no single worldwide standard except for the seven epochs making up the last two periods.

Over the interval from the Paleozoic to the present, about 38 epochs are recognized. This interval is represented by approximately 250 formations, discrete layers thick enough and distinctive enough in lithology to merit delineation as units of the geologic column. Also employed in subdivision is the zone concept, in which it is the fossils in the rocks rather than the lithologic character that defines minor stratigraphic boundaries. The basis of zone definition varies among geologists, some considering a zone to be all rocks containing a certain species (usually an invertebrate). Other geologists focus on special fossil assemblages.

The lower part of the geologic column, where fossils are very scarce, was at one time viewed in the context of two eras of time, but subsequent mapping has shown the provincial bias in such a scheme. Consequently, the entire lower column is now considered a single unit, the Precambrian. The results of isotopic dating are now providing finer Precambrian subdivisions that have worldwide applicability.

The geologic column and the relative geologic time scale are sufficiently defined to fulfill the use originally envisioned for them—providing a framework within which to tell the story of Earth history. Just as human history has its interweaving plots of warfare, cultural development, and technological advance, so the Earth's rocks tell another story of intertwined sequences of events. Mountains have been built and eroded away, seas have advanced and retreated, a myriad of life-forms has inhabited land and sea. In all these happenings, the geologic column and its associated time scale spell the difference between an unordered series of isolated events and the unfolding story of a changing Earth.

ABSOLUTE DATING

Although relative ages can generally be established on a local scale, the events recorded in rocks from different locations can be integrated into a picture of regional or global scale only if their sequence in time is firmly established. The time that has elapsed since certain minerals formed can now be determined because of the presence of a small amount of natural radioactive atoms in their structures. Whereas studies using fossil dating began almost 300 years ago, radioactivity itself was not discovered until roughly a century ago, and it has only been from about 1950 that extensive efforts to date geologic materials have become common. Methods of isotopic measurement continue to be refined today, and absolute dating has become an essential component of virtually all field-oriented geologic investigations. In the process of refining isotopic measurements, methods for low-contamination chemistry had to be developed, and it is significant that many such methods now in worldwide use resulted directly from work in geochronology.

It has already been explained how different Earth processes create different rocks as part of what can be considered a giant rock-forming and rock-re-forming cycle. Attention has been called wherever possible to those rocks that contain minerals suitable for precise isotopic dating. It is important to remember that precise ages cannot be obtained for just any rock unit but that any unit can be dated relative to a datable unit. The following discussion will show why this is so, treating in some detail the analytic and geologic problems that have to be overcome if precise ages are to be determined. It will become apparent, for example, that isotopic ages can be reset by high temperatures. However, this seeming disadvantage can be turned to one's favour in determining the cooling history of a rock. As various dating methods are discussed, the great interdependence of the geologic and analytic components essential to geochronology should become evident.

The field of isotope geology complements geochronology. Workers in isotope geology follow the migration of isotopes produced by radioactive decay through large- and small-scale geologic processes. Isotopic tracers of this kind can be thought of as an invisible dye injected by nature into Earth systems that can be observed only with sophisticated instruments. Studying the movement or distribution of these isotopes can provide insights into the nature of geologic processes.

The Principles of Isotopic Dating

All absolute isotopic ages are based on radioactive decay, a process whereby a specific atom or isotope is converted into another specific atom or isotope at a constant and known rate. Most elements exist in different atomic forms that are identical in their chemical properties but differ in the number of neutral particles (neutrons) in the nucleus. For a single element, these atoms are called isotopes. Because isotopes differ in mass, their relative abundance can be determined if the masses are separated in a mass spectrometer.

Radioactive decay can be observed in the laboratory by either of two means: (1) a radiation counter (such as a Geiger counter), which detects the number of high-energy particles emitted by the disintegration of radioactive atoms in a sample of geologic material, or (2) a mass spectrometer, which permits the identification of daughter atoms formed by the decay process in a sample containing radioactive parent atoms. The particles given off during the decay process are part of a profound fundamental change in the nucleus. To compensate for the loss of mass (and energy), the radioactive atom undergoes internal transformation and in most cases simply becomes an atom of a different chemical element. In terms of the numbers of atoms present, it is as if apples changed spontaneously into oranges at a fixed and known rate. In this analogy, the apples would represent radioactive, or parent, atoms, while the oranges would represent the atoms formed, the so-called daughters. Pursuing this analogy further, one would expect that a new basket of apples would have no oranges but that an older one would have many. In fact, one would expect that the ratio of oranges to apples would change in a very specific way over the time elapsed, since the process continues until all the apples are converted. In geochronology the situation is identical. A particular rock or mineral that contains a radioactive isotope (or radioisotope) is analyzed to determine the number of parent and daughter isotopes present, whereby the time since that mineral or rock formed is calculated. Of course, one must select geologic materials that contain elements with long half-lives—that is, those for which some parent atoms would remain.

Given below is the simple mathematical relationship that allows the time elapsed to be calculated from the measured parent/daughter ratio. The age calculated is only as good as the existing knowledge of the decay rate and is valid only if this rate is constant over the time that elapsed.

Fortunately for geochronology the study of radioactivity has been the subject of extensive theoretical and laboratory investigation by physicists for almost a century. The results show that there is no known process that can alter the rate of radioactive decay. By way of explanation, it can be noted that since the cause of the process lies deep within the atomic nucleus, external forces such as extreme heat and pressure have no effect. The same is true regarding gravitational, magnetic, and electric fields, as well as the chemical state in which the atom resides. In short, the process of radioactive decay is immutable under all known conditions. Although it is impossible to predict when a particular atom will change, given a sufficient number of atoms, the rate of their decay is found to be constant. The situation is analogous to the death rate among human populations insured by an insurance company. Even though it is impossible to predict when a given policyholder will die, the company can count on paying off a certain number of beneficiaries every month. The recognition that the rate of decay of any radioactive parent atom is proportional to the number of atoms (N) of the parent remaining at any time gives rise to the following expression:

R	$\propto$	N	
rate of disintegration	is proportional to	number of parent atoms present.	(1)

Converting this proportion to an equation incorporates the additional observation that different radioisotopes have different disintegration rates even when the same number of atoms are observed undergoing decay. In other words, each radioisotope has its own decay constant, abbreviated λ, which provides a measure of its intrinsic rapidity of decay. Proportion 1 becomes:

$$R = \lambda N. \qquad (2)$$

Stated in words, this equation says that the rate at which a certain radioisotope disintegrates depends not only on how many atoms of that isotope are present but also on an intrinsic property of that isotope represented by λ, the so-called decay constant. Values of λ vary widely—from 10^{20} reciprocal seconds (that is, the unit of 1 second) for a rapidly disintegrating isotope such as helium-5 to less than 10^{-25} reciprocal seconds for slowly decaying cerium-142.

In the calculus, the rate of decay R in equation 2 is written as the derivative dN/dt, in which dN represents the small number of atoms that decay in an infinitesimally short time interval dt. Replacing R by its equivalent dN/dt results in the differential equation

$$\frac{dN}{dt} = -\lambda N. \qquad (3)$$

Solution of this equation by techniques of the calculus yields one form of the fundamental equation for radiometric age determination,

$$\frac{N}{N_0} = e^{-\lambda t}, \qquad (4)$$

in which N_o is the number of radioactive atoms present in a sample at time zero, N is the number of radioactive atoms present in the sample today, e is the base of natural logarithms (equal to about 2.72), λ is the decay constant of the radioisotope being considered, and t is the time elapsed since time zero.

Two alterations are generally made to equation 4 in order to obtain the form most useful for radiometric dating. In the first place, since the unknown term in radiometric dating is obviously t, it is desirable to rearrange equation 4 so that it is explicitly solved for t. Second, the more common way to express the intrinsic decay rate of a radioisotope is through its half-life (abbreviated $t_{1/2}$) rather than through the decay constant λ. Half-life is defined as the time period that must elapse in order to halve the initial number of radioactive atoms. The half-life and the decay constant are inversely proportional because rapidly decaying radioisotopes have a high decay constant but a short half-life. With t made explicit and half-life introduced, equation 4 is converted to the following form, in which the symbols have the same meaning:

$$t = \frac{t_{1/2}}{0.693} \times \log_e\left(\frac{N_0}{N}\right). \qquad (5)$$

Alternatively, because the number of daughter atoms is directly observed rather than N, which is the initial number of parent atoms present, another formulation may be more convenient. Since the initial number of parent atoms present at time zero N_o must be the sum of the parent atoms remaining N and the daughter atoms present D, one can write:

$$D = N_0 - N. \qquad (6)$$

From equation 4 above, it follows that $N_o = N(e^{\lambda t})$. Substituting this in equation 6 gives

$$D = Ne^{\lambda t} - N,$$

$$\text{or} \quad D = N(e^{\lambda t} - 1),$$

$$\text{or} \quad \frac{D}{N} = (e^{\lambda t} - 1).$$

If one chooses to use P to designate the parent atom, the expression assumes its familiar form:

$$\frac{D}{P} = (e^{\lambda t} - 1) \qquad (7)$$

and

$$t = \frac{1}{\lambda} \ln\left(\frac{D}{P} + 1\right). \qquad (8)$$

This pair of equations states rigorously what might be assumed from intuition, that minerals formed at successively longer times in the past would have progressively higher daughter-to-parent ratios. This follows because, as each parent atom loses its identity with time, it reappears as a daughter atom. The increase in D/P with time is evident in equation (7) because larger values of time will increase the value of $e^{\lambda t}$, where λ is constant. Equation (8) documents the simplicity of direct isotopic dating. The time of decay is proportional to the natural logarithm (represented by ln) of the ratio of D to P. In short, one need only measure the ratio of the number of radioactive parent and daughter atoms present, and the time elapsed since the mineral or rock formed can be calculated, provided of course that the decay rate is known. Likewise, the conditions that must be met to make the calculated age precise and meaningful are in themselves simple:

1. The rock or mineral must have remained closed to the addition or escape of parent and daughter atoms since the time that the rock or mineral (system) formed.
2. It must be possible to correct for other atoms identical to daughter atoms already present when the rock or mineral formed.
3. The decay constant must be known.
4. The measurement of the daughter-to-parent ratio must be accurate because uncertainty in this ratio contributes directly to uncertainty in the age.

Different schemes have been developed to deal with the critical assumptions stated above. In uranium-lead dating, minerals virtually free of initial lead can be isolated and corrections made for the trivial amounts present. In whole rock isochron methods that make use of the rubidium-strontium or samarium-neodymium decay schemes, a series of rocks or minerals are chosen that can be assumed to have the same age and identical abundances of their initial isotopic ratios. The results are then tested for the internal consistency that can validate the assumptions. In all cases, it is the obligation of the investigator making the determinations to include enough tests to indicate that the absolute age quoted is valid within the limits stated. In other words, it is the obligation of geochronologists to try to prove themselves wrong by including a series of cross-checks in their measurements before they publish a result. Such checks include dating a series of ancient units with closely spaced but known relative ages and replicate analysis of different parts of the same rock body with samples collected at widely spaced localities.

The importance of internal checks as well as interlaboratory comparisons becomes all the more apparent when one realizes that geochronology laboratories are limited in number. Because of the expensive equipment necessary and the combination of geologic, chemical, and laboratory skills required, geochronology is usually carried out by teams of experts. Most geologists must rely on geochronologists for their results. In turn, the geochronologist relies on the geologist for relative ages.

The Origin of Radioactive Elements used in Dating

In order for a radioactive parent-daughter pair to be useful for dating, many criteria must be met. This section examines these criteria and explores the ways in which the reliability of the ages measured can be assessed. Because geologic materials are diverse in their origin and chemical content and datable elements are unequally distributed, each method has its strengths and weaknesses.

When the elements in the Earth were first created, many radioactive isotopes were present. Of these, only the radioisotopes with extremely long half-lives remain. It should be mentioned in passing that some of the radioisotopes present early in the history of the solar system and now completely extinct have been recorded in meteorites in the form of the elevated abundances of their daughter isotopes. Analysis of such meteorites makes it possible to estimate the time that elapsed between element creation and meteorite formation. Natural elements that are still radioactive today produce daughter products at a very slow rate. Hence, it is easy to date very old minerals but difficult to obtain the age of those formed in the recent geologic past. This follows from the fact that the amount of daughter isotopes present is so small that it is difficult to measure. The difficulty can be overcome to some degree by achieving lower background contamination, by improving instrument sensitivity, and by finding minerals with abundant parent isotopes. Geologic events of the not-too-distant past are more easily dated by using recently formed radioisotopes with short half-lives that produce more daughter products per unit time. Two sources of such isotopes exist. In one case, intermediate isotopes in the uranium or thorium decay chain can become isolated in certain minerals due to differences in chemical properties and, once fixed, can decay to new isotopes, providing a measure of the time elapsed since they were isolated. To understand this, one needs to know that though uranium-238 (^{238}U) does indeed decay to lead-206 (^{206}Pb), it is not a one-step process. In fact, this is a multistep process involving the expulsion of eight alpha particles and six beta particles, along with a considerable amount of energy. There exists a series of different elements, each of them in a steady state where they form at the same rate as they disintegrate. The number present is proportional to their decay rate, with long-lived members being more abundant. Because all of these isotopes have relatively short half-lives, none remains since the creation of the elements, but instead they are continuously provided by the decay of the long-lived parent. This type of dating, known as disequilibrium dating, will be explored below in the section Uranium-series disequilibrium dating.

Another special type of dating employs recently formed radioisotopes produced by cosmic-ray bombardment of target atoms at the Earth's surface or in the atmosphere. The amounts produced, although small, provide insight into many near-surface processes in the geologic past. This aspect of geology is becoming increasingly important as researchers try to read the global changes that took place during the Earth's recent past in an effort to understand or predict the future. The most widely used radioactive cosmogenic isotope is carbon of mass 14 (^{14}C), which provides a method of dating events that have occurred over roughly the past 50,000 years. This time spans much of the historic and prehistoric record of mankind.

The Isochron Method

Many radioactive dating methods are based on minute additions of daughter products to a rock or mineral in which a considerable amount of daughter-type isotopes already exists. These isotopes did not come from radioactive decay in the system but rather formed during the original creation of the elements. In this case, it is a big advantage to present the data in a form in which the abundance of both the parent and daughter isotopes are given with respect to the abundance of the initial background daughter. The incremental additions of the daughter type can then be viewed in proportion to the abundance of parent atoms. In mathematical terms this is achieved as follows. It has already been shown—equation 7—that the number of daughter atoms present from radioactive decay D^* can be related to the number of parent atoms remaining P by the simple expression:

$$D^* = P(e^{\lambda t} - 1). \qquad (9)$$

When some daughter atoms are initially present (designated D_o), the total number D is the sum of radiogenic and initial atoms, so that

$$D = D_0 + P(e^{\lambda t} - 1). \qquad (10)$$

To establish the condition that both parent and daughter abundances should be relative to the initial background, a stable isotope S of the daughter element can be chosen and divided into all portions of this equation; thus,

$$\frac{D}{S} = \left(\frac{D}{S}\right)_0 + \frac{P}{S}(e^{\lambda t} - 1).$$

This equation has the form; $y = b + xm$, which is that of a straight line on x–y coordinates. The slope m is equal to $(e^{\lambda t} - 1)$ and the intercept is equal to $(D/S)_o$. This term is called the initial ratio. The slope is proportional to the geologic age of the system.

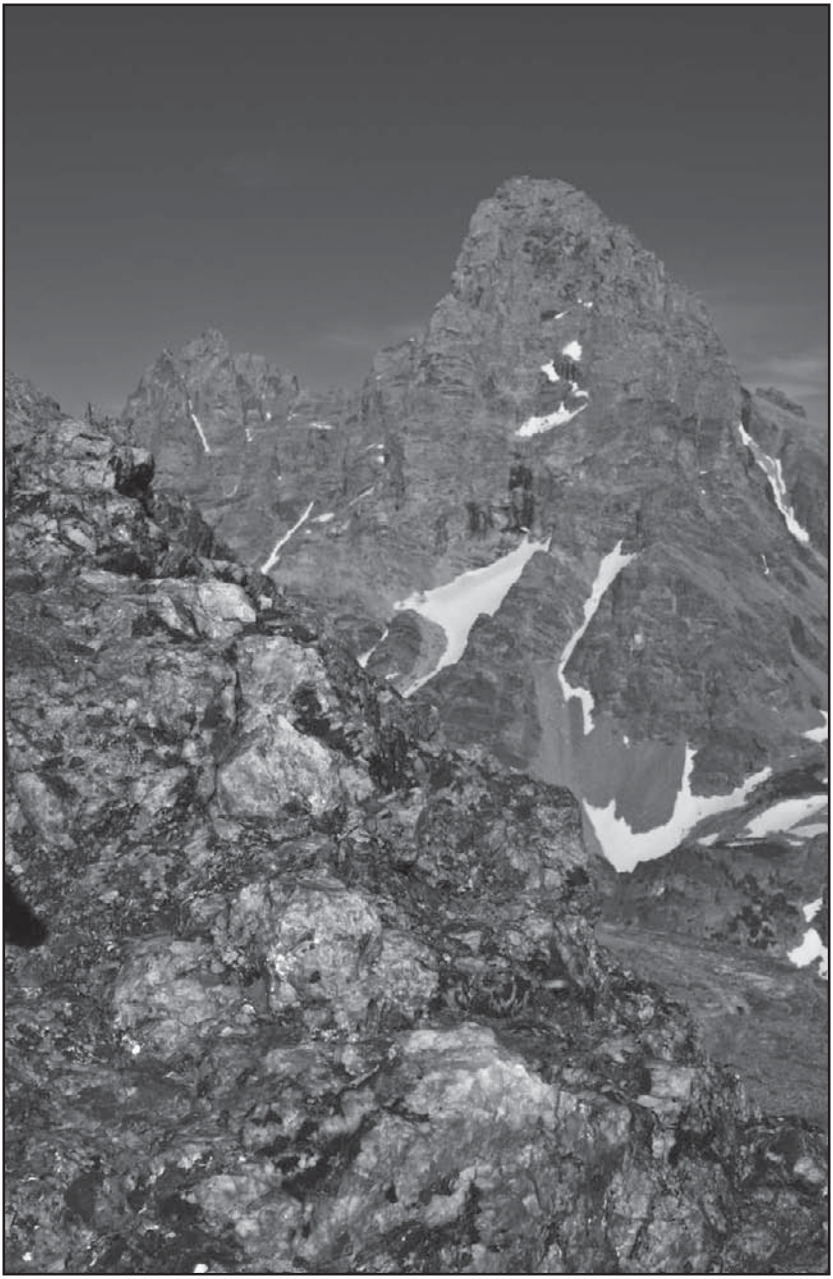
In practice, the isochron approach has many inherent advantages. When a single body of liquid rock crystallizes, parent and daughter elements may separate so that, once solid, the isotopic data would define a series of points, such as those shown as open circles designated R_1, R_2, R_3 in the figure below.

They plot along a horizontal line reflecting a common value for the initial daughter isotope ratio $(D/S)_0$. With time each would then develop additional daughter abundances in proportion to the amount of parent present. If a number of samples are analyzed and the results are shown to define a straight line within error, then a precise age is defined because this is only possible if each is a closed system and each has the same initial ratio and age. The uncertainty in determining the slope is reduced because it is defined by many points. A second advantage of the method relates to the fact that under high-temperature conditions the daughter isotopes may escape from the host minerals. In this case, a valid age can still be obtained, provided that they remain within the rock. Should a point plot below the line, it could indicate that a particular sample was open to migration of the dating elements or that the sample was contaminated and lay below the isochron when the rock solidified.

Rubidium-strontium (Rb-Sr) dating was the first technique in which the whole rock isochron method was extensively employed. Certain rocks that cooled quickly at the surface were found to give precisely defined linear isochrons, but many others did not. Some studies have shown that rubidium is very mobile both in fluids that migrate through the rock as it cools and in fluids that are present as the rock undergoes chemical weathering. Similar studies have shown that the samarium-neodymium (Sm-Nd) parent-daughter pair is more resistant to secondary migration but that, in this instance, sufficient initial spread in the abundance of the parent isotope is difficult to achieve.

The Analysis of Separated Minerals

When an igneous rock crystallizes, a wide variety of major and trace minerals may form, each concentrating certain elements and radioactive trace elements within the rock. By careful selection, certain minerals that contain little or no daughter element but abundant parent element can be analyzed. In this case, the slope of the line in the figure is computed from an assumed value for the initial ratio, and it is usually possible to show that uncertainties related to this assumption are negligible. This is possible in potassium-argon (K-Ar) dating, for example, because most minerals do not take argon into their structures initially. In rubidium-strontium dating, micas exclude strontium when they form, but accept much rubidium. In uranium-lead (U-Pb) dating of zircon, the zircon is found to exclude initial lead almost completely. Minerals, too, are predictable chemical compounds that can be shown to form at specific temperatures and remain closed up to certain temperatures if a rock has been reheated or altered. A rock, on the other hand, may contain minerals formed at more than one time under a variety of conditions. Under such circumstances the isolation and analysis of certain minerals can indicate at what time these conditions prevailed. If a simple mineral is widespread in the geologic record, it is more valuable for dating as more units can be measured for age and compared by the same method. However, if a single parent-daughter pair that is amenable to precise analysis can be measured in a variety of minerals, the ages of a wide variety of rock types can be determined by a single method without the need for intercalibration. In some cases the discovery of a rare trace mineral results in a major breakthrough as it allows precise ages to be determined in formerly undatable units.



Modern scientific techniques help scientists measure the age of ancient rocks such as these pegmatite with quartz and feldspar ones in Wyoming's Teton Range. Dr. Marli Miller/Visuals Unlimited/Getty Images

For example, the mineral baddeleyite, an oxide of zirconium (ZrO_2), has been shown to be widespread in small amounts in mafic igneous rocks (that is, those composed primarily of one or more ferromagnesian, dark-coloured minerals). Here, a single uranium-lead isotopic analysis can provide an age more precise than can be obtained by the whole rock isochron method involving many analyses. When single minerals are analyzed, each grain can be studied under a microscope under intense side light so that alterations or imperfections can be revealed and excluded. If minerals are used for dating, the necessary checks on the ages are achieved by analyzing samples from more than one location and by analyzing different grain sizes or mineral types that respond differently to disturbing events. It can be said that minerals provide a high degree of sample integrity that can be predicted on the basis of experience gained through numerous investigations under a variety of geologic conditions. An ideal mineral is one that has sufficient parent and daughter isotopes to measure precisely, is chemically inert, contains little or no significant initial daughter isotopes, and retains daughter products at the highest possible temperatures. A specific datable mineral like rutile, which can be linked to a specific event such as the formation of a mineral deposit, is especially important.

Model Ages

Since the Earth was formed, the abundance of daughter product isotopes has increased through time. For example, the ratio of lead of mass 206 relative to that of mass 204 has changed from an initial value of about 10 present when the Earth was formed to an average value of about 19 in rocks at the terrestrial surface today. This is true because uranium is continuously creating more lead. A lead-rich mineral formed and isolated early in Earth history would have a low lead-206 to lead-204 ratio because it did not receive subsequent additions by the radioactive decay of uranium. If the Earth's interior were a simple and homogeneous reservoir with respect to the ratio of uranium to lead, a single sample extracted by a volcano would provide the time of extraction. This would be called a model age. No parent-daughter value for a closed system is involved, rather just a single isotopic measurement of lead viewed with respect to the expected evolution of lead in the Earth. Unfortunately the simplifying assumption in this case is not true, and lead model ages are approximate at best. Other model ages can be calculated using neodymium isotopes by extrapolating present values back to a proposed mantle-evolution line. In both cases, approximate ages that have a degree of validity with respect to one another result, but they are progressively less reliable as the assumptions on which the model is calculated are violated.

The progressive increase in the abundance of daughter isotopes over time gains a special significance where the parent element is preferentially enriched in either the mantle or the crust. For example, rubidium is concentrated in the crust, and as a result the present-day continents, subjected to weathering, have an elevated radiogenic to stable isotope ratio ($^{87}Sr/^{86}Sr$) of 0.720. In contrast, modern volcanic rocks in the oceans imply that much of the mantle has a value between about 0.703 and 0.705. Should crustal material be recycled, the strontium isotopic signature of the melt would be diagnostic.

Multiple Ages for a Single Rock; the Thermal Effect

Fossils record the initial, or primary, age of a rock unit. Isotopic systems, on the other hand, can yield either the primary age or the time of a later event, because crystalline materials are very specific in the types of atoms they incorporate, in terms of both the atomic size and charge. An element formed by radioactive decay is quite different from its parent atom and thus is out of place with respect to the host mineral. All it takes for such an element to be purged from the mineral is sufficient heat to allow solid diffusion to occur. Each mineral has a temperature at which rapid diffusion sets in, so that, as a region is slowly heated, first one mineral and then another loses its daughter isotopes. When this happens, the isotopic "clock" is reset to zero, where it remains until the mineral cools below the blocking temperature. (This is the temperature below which a mineral becomes a closed chemical system for a specific radioactive decay series. Accordingly, the parent-daughter isotope ratio indicates the time elapsed since that critical threshold was reached.) In this case, the host mineral could have an absolute age very much older than is recorded in the isotopic record. The isotopic age then is called a cooling age. It is even possible by using a series of minerals with different blocking temperatures to establish a cooling history of a rock body—that is, the times since the rock body cooled below successively lower temperatures. Such attempts can be complicated by the fact that a mineral may "grow" below the blocking temperature rather than simply become closed to isotopic migration. When this happens, the age has little to do with the cooling time. Another problem arises if a region undergoes a second reheating event. Certain minerals may record the first event, whereas others may record the second, and any suggestion of progressive cooling between the two is invalid. This complication does not arise when rapid cooling has occurred. Identical ages for a variety of minerals with widely different blocking temperatures is unequivocal proof of rapid cooling.

Fortunately for geologists the rock itself records in its texture and mineral content the conditions of its formation. A rock formed at the surface with no indication of deep burial or new mineral growth can be expected to give a valid primary age by virtue of minerals with low blocking temperatures. On the other hand, low-blocking-point minerals from a rock containing minerals indicative of high temperatures and pressures cannot give a valid primary age. Such minerals would be expected to remain open until deep-level rocks of this sort were uplifted and cooled.

Given these complicating factors, one can readily understand why geochronologists spend a great deal of their time and effort trying to see through thermal events that occurred after a rock formed. The importance of identifying and analyzing minerals with high blocking temperatures also cannot be overstated. Minerals with high blocking temperatures that form only at high temperatures are especially valuable. Once formed, these minerals can resist daughter loss and record the primary age even though they remained hot (say, 700° C [1,292° F]) for a long time. The mineral zircon datable by the uranium-lead method is one such mineral. The mica mineral biotite dated by either the potassium-argon or the rubidium-strontium method occupies the opposite end of the spectrum and does not retain daughter products until cooled below about 300° C (572° F). Successively higher blocking temperatures are recorded for another mica type known as muscovite and for amphibole, but the ages of both of these minerals can be completely reset at temperatures that have little or no effect on zircon.

Taken in perspective, it is evident that many parts of the Earth's crust have experienced reheating temperatures above 300° C (572° F). For example, reset mica ages are very common in rocks formed at deep crustal levels. Vast areas within the Precambrian shield, which have identical ages reflecting a common cooling history, have been identified. These are called geologic provinces. By contrast, rocks that have approached their melting point, say, 750° C (1,382° F)—which can cause new zircon growth during a second thermal event—are rare, and those that have done this more than once are almost nonexistent.

Instruments and Procedures

The process of isotopic dating was improved immensely in 1950 with the development of mass spectrometers capable of measuring very tiny changes in mass. Since then a number of advanced instruments and techniques have been developed to discern the age of a rock or mineral sample. Such advances include the processes of isotope dilution and ion-exchange chromatography and the development of accelerator and thermal ionization mass spectrometers.

The Use of Mass Spectrometers

The age of a geologic sample is measured on as little as a billionth of a gram of daughter isotopes. Moreover, all the isotopes of a given chemical element are nearly identical except for a very small difference in mass. Such conditions necessitate instrumentation of high precision and sensitivity. Both these requirements are met by the modern mass spectrometer. A high-resolution mass spectrometer of the type used today was first described by the American physicist Alfred O. Nier in 1940, but it was not until about 1950 that such instruments became available for geochronological research.

For isotopic dating with a mass spectrometer, a beam of charged atoms, or ions, of a single element from the sample is produced. This beam is passed through a strong magnetic field in a vacuum, where it is separated into a number of beams, each containing atoms of only the same mass. Because of the unit electric charge on every atom, the number of atoms in each beam can be evaluated by collecting individual beams sequentially in a device called a Faraday cup. Once in this collector, the current carried by the atoms is measured as it leaks across a resistor to ground. Currents measured are small, only from 10^{-11} to 10^{-15} ampere, so that shielding and preamplification are required as close to the Faraday cup as possible. It is not possible simply to count the atoms, because all atoms loaded into the source do not form ions and some ions are lost in transmission down the flight tube. Precise and accurate information as to the number of atoms in the sample can, however, be obtained by measuring the ratio of the number of atoms in the various separated beams. By adding a special artificially enriched isotope during sample dissolution and by measuring the ratio of natural to enriched isotopes in adjacent beams, the number of daughter isotopes can be readily determined. The artificially enriched isotope is called a "spike." It is usually a highly purified form of a low-abundance natural isotope, but an even better spike is an isotope with a mass not found in nature at all. Lead-205 produced in a type of particle accelerator called a cyclotron constitutes such an ideal spike.

As the sample is heated and vaporizes under the vacuum in the source area of the mass spectrometer, it is commonly observed that the lighter isotopes come off first, causing a bias in the measured values that changes during the analysis. In most cases this bias, or fractionation, can be corrected if the precise ratio of two of the stable isotopes present is known. Today's state-of-the-art instruments produce values for strontium and neodymium isotopic abundances that are reproducible at a level of about 1 in 20,000. Such precision is often essential in the isochron method because of the small changes in relative daughter abundance that occur over geologic time.

Technical Advances

The ability to add a single artificial mass to the spectrum in a known amount and to determine the abundances of other isotopes with respect to this provides a powerful analytical tool. By means of this process known as isotope dilution, invisibly small amounts of material can be analyzed, and because only ratios are involved, a loss of part of the sample during preparation has no effect on the result. Spike solutions can be calibrated simply by obtaining a highly purified form of the element being calibrated. After carefully removing surface contamination, a precisely weighted portion of the element is dissolved in highly purified acid and diluted to the desired level in a weighed quantity of water. What is required is dilution of one cubic centimetre to a litre from which a second cubic centimetre is again diluted to a litre to approach the range of parts per million or parts per billion typically encountered in samples. In this way, a known number of natural isotopes can be mixed with a known amount of spike and the concentration in the spike solution determined from the ratio of the masses. Once the calibration has been completed, the process is reversed and a weighed amount of spike is mixed with the parent and daughter elements from a mineral or rock. The ratio of the masses then gives the number of naturally produced atoms in the sample. The use of calibrated enriched isotopic tracers facilitates checks for contamination, even though the process is time-consuming. A small but known amount of tracer added to a beaker of water can be evaporated under clean-room conditions. Once loaded in a mass spectrometer, the contamination from the beaker and the water is easily assessed with respect to the amount of spike added. Contamination as small as 10^{-12} gram can be detected by this method.

The materials analyzed during isotopic investigations vary from microgram quantities of highly purified mineral grains to gram-sized quantities of rock powders. In all cases, the material must be dissolved without significant contamination. The spike should be added before dissolution. Most of the minerals in rocks can be dissolved in a day or so at a temperature near 100° C (212° F). Certain minerals that are highly refractory both in nature and in the laboratory (zircon, for example) may require five days or more at temperatures near 220° C (428° F). In this case, the sample is confined in a solid Teflon (trade name for a synthetic resin composed of polytetrafluoroethylene), metal-clad pressure vessel, introduced by the Canadian geochronologist Thomas E. Krogh in 1973.

The method just described proved to be a major technical breakthrough as it resulted in a reduction in lead-background contamination by a factor of between 10,000 and nearly 1,000,000. This means that a single grain can now be analyzed with a lower contamination level (or background correction) than was possible before with 100,000 similar grains. Advances in high-sensitivity mass spectrometry of course were essential to this development.

Once dissolved, the sample is ready for the chemical separation of the dating elements. This is generally achieved by using the methods of ion-exchange chromatography. In this process, ions are variously adsorbed from solution onto materials with ionic charges on their surface and separated from the rest of the sample. After the dating elements have been isolated, they are loaded into a mass spectrometer and their relative isotopic abundances determined.

The abundance of certain isotopes used for dating is determined by counting the number of disintegrations per minute (that is, emission activity). The rate is related to the number of such atoms present through the half-life. For example, a certain amount of carbon-14 (^{14}C) is present in all biological components at the Earth's surface. This radioactive carbon is continually formed when nitrogen atoms of the upper atmosphere collide with neutrons produced by the interaction of high-energy cosmic rays with the atmosphere. An organism takes in small amounts of carbon-14, together with the stable (nonradioactive) isotopes carbon-12 (^{12}C) and carbon-13 (^{13}C), as long as it is alive. Once it dies, however, no additional carbon-14 is acquired and the level of radiocarbon in the organism's tissue decreases progressively as a function of half-life. The time that has passed since the organism was alive can be determined by counting the beta emissions from a tissue sample. The number of emissions in a given time period is proportional to the amount of residual carbon-14.

The introduction of an instrument called an accelerator mass spectrometer has brought about a major advance in radiocarbon dating. Unlike the old detector (such as the Geiger counter), which counts the few decay particles emitted from a large amount of carbon, the new instrument counts directly all of the carbon-14 atoms in a sample. This increase in instrument sensitivity has made it possible to reduce the sample size by as much as 10,000 times and at the same time improve the precision of ages measured.

In a similar development, the use of highly sensitive thermal ionization mass spectrometers is replacing the counting techniques employed in some disequilibrium dating. Not only has this led to a reduction in sample size and measurement errors, but it also has permitted a whole new range of problems to be investigated. Certain parent-daughter isotopes are extremely refractory and do not ionize in a conventional mass spectrometer. To solve this problem, researchers are developing new instruments in which a small amount of material can be evaporated from the surface with a pulse of energy and ionized with a pulse of laser light. A major trend anticipated in geochronology and isotope geochemistry involves the analysis of mineral grains in place without chemical dissolution and mass spectrometry. This type of analysis requires expensive equipment in which a focused beam of ions is directed at a spot on a mineral sample. This causes atoms to evaporate from the surface, and the ions produced are extracted and measured in a mass spectrometer. Uranium-lead dating of zircon by this method has been pioneered by William Compston at the Australian National University.

The Major Methods of Isotopic Dating

Isotopic dating relative to fossil dating requires a great deal of effort and depends on the integrated specialized skills of geologists, chemists, and physicists. It is, nevertheless, a valuable resource that allows correlations to be made over virtually all of Earth history with a precision once only possible with fossiliferous units that are restricted to the most recent 12 percent or so of geologic time. Although any method may be attempted on any unit, the best use of this resource requires that every effort be made to tackle each problem with the most efficient technique. Because of the long half-life of some isotopic systems or the high background or restricted range of parent abundances, some methods are inherently more precise. The skill of a geochronologist is demonstrated by the ability to attain the knowledge required and the precision necessary with the least number of analyses. The factors considered in selecting a particular approach are explored here.

The Uranium-Lead Method

As each dating method was developed, tested, and improved—mainly since 1950—a vast body of knowledge about the behaviour of different isotopic systems under different geologic conditions has evolved. It is now clear that with recent advances the uranium-lead method is superior in providing precise age information with the least number of assumptions. The method has evolved mainly around the mineral zircon ($ZrSiO_4$). Because of the limited occurrence of this mineral, it was once true that only certain felsic igneous rocks (those consisting largely of the light-coloured, silicon and aluminum-rich minerals feldspar and quartz) could be dated. Today, however, baddeleyite (ZrO_2) has been found to be widespread in the silica-poor mafic igneous rocks. In addition, perovskite ($CaTiO_3$), a common constituent of some ultramafic igneous rocks, has been shown to be amenable to precise uranium-lead dating. As a result of these developments, virtually all igneous rocks can now be dated. This capability, moreover, has been enhanced because the most advanced geochronological laboratories are able to analyze samples that weigh only a few millionths of a gram. This amount can be found in a comparatively large number of rocks, whereas the amount previously required (about 0.1 gram) cannot. Age determinations also can now be made of low-uranium trace minerals such as rutile (TiO_2), a common constituent found in mineral deposits, adding still further to the number of entities that are datable by the uranium-lead method. Other minerals commonly employed to date igneous and metamorphic rocks include titanite, monazite, and even garnet in certain favourable cases. Additional minerals have been tried with varying success.

Double Uranium-Lead Chronometers

The reason why uranium-lead dating is superior to other methods is simple: there are two uranium-lead chronometers. Because there exist two radioactive uranium atoms (those of mass 235 and 238), two uranium-lead ages can be calculated for every analysis. The age results or equivalent daughter-parent ratio can then be plotted one against the other on a concordia diagram. If the point falls on the upper curve shown, the locus of identical ages, the result is said to be concordant, and a closed-system unequivocal age has been established. Any leakage of daughter isotopes from the system will cause the two ages calculated to differ, and data will plot below the curve. Because each of the daughters has a different half-life, early leakage will affect one system more than the other. Thus, there is a built-in mechanism that can prove or disprove whether a valid age has been measured. Historically it had been observed that the uranium-lead systems in the mineral zircon from unmetamorphosed rocks were almost invariably disturbed or discordant but yielded a linear array on the concordia diagram. Given a set of variably disturbed samples, an extrapolation to zero disturbance was possible. More recently, it has been found that of all the grains present in a rock a very few still retain closed isotopic systems but only in their interior parts. Thus grains with a diameter comparable to that of a human hair, selected under a microscope to be crack-free and of the highest possible quality, have been found to be more concordant than cracked grains. In addition, it has been shown that most such grains can be made much more concordant by mechanically removing their outer parts using an air-abrasion technique (upper points in). Of course, the ability to analyze samples weighing only a few millionths of a gram was essential to this development. As noted earlier, this in turn was possible solely because the lead background contamination had been reduced from 1×10^{-6} grams to almost 1×10^{-12} grams per analysis. The methods of selection and abrasion used to locate grains with closed isotopic systems could be worked out only because the uranium-lead method has the inherent ability to assess with a single analysis whether or not a closed isotopic system has prevailed.

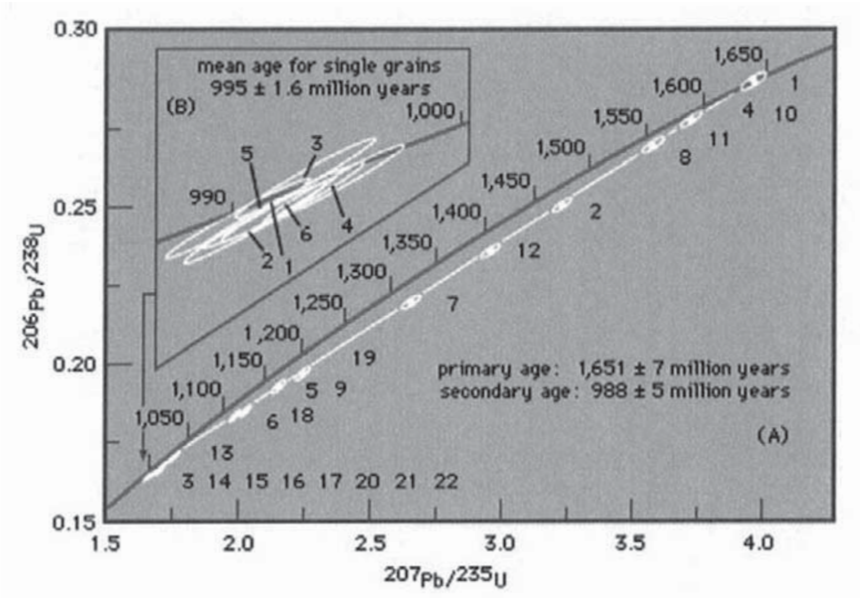
The presence of two radioactive parents provides a second major advantage because, as daughter products, lead atoms are formed at different rates and their relative abundance undergoes large changes as a function of time. Thus, the ratio of lead-207 to lead-206 changes by about 0.1 percent every two million years. Since this ratio is easily calibrated and reproduced at such a level of precision, errors as low as ±2 million years at a confidence level of 95 percent are routinely obtained on lead-207–lead-206 ages. By contrast, errors as high as ±30 to 50 million years are usually quoted for the rubidium-strontium and samarium-neodymium isochron methods.

The Importance of Zircon in Uranium-Lead Dating

The mineral zircon adds three more fundamental advantages to uranium-lead dating. First, its crystal structure allows a small amount of tetravalent uranium to substitute for zirconium, but excludes with great efficiency the incorporation of lead. (It might be said that one begins with an empty box.) Second, zircon, once formed, is highly resistant to change and has the highest blocking temperature ever observed. Finally, with few predictable exceptions, zircon grows or regrows only in liquid rock or in solid rock reheated to approach its melting point. Combining all of these attributes, it is often possible to measure both the time of crystallization and the time of second melting in different parts of the same grain or in different selected grains from the same rock. Of course, such a high blocking temperature can have its disadvantages. Inherited cores may give a mixed false age when the age of crystallization is sought. For this reason, three or more grain types or parts of a grain are analyzed to establish that material of only one age is present.

Experience with the results of the uranium-lead method for zircons has demonstrated an interesting paradox. If left at low surface temperatures for a geologically long time, the radioactivity within the crystal can destroy the crystal lattice structure, whereas at higher temperatures this process is self-annealing. In fact, when examined by X-ray methods, some zircons have no detectable structure, indicating that at least 25 percent of the initial atoms have been displaced by radiation damage. Under these conditions a low-temperature event insufficient to even reset the potassium-argon system in biotite can cause lead to be lost in some grains. It is no coincidence that, when criteria were finally found to locate concordant grains, these grains were also found to be those with the lowest uranium content and the lowest related radiation damage.

Given the two related uranium-lead parent-daughter systems, it is possible to determine both the time of the initial, or primary, rock-forming event and the time of a major reheating, or secondary, event. For example, the uranium-lead isotopes in the mineral titanite ($CaTiSiO_5$) from a series of rocks that have a common geologic history can be plotted on a straight line (as shown in the figure on page 125). The minerals first formed 1,651 million years ago but were later heated and lost varying amounts of lead 986 million years ago. In many cases, new titanite, distinguishable on the basis of colour, has formed in the same rock, while older, partly reset titanite is still present. Data points 14 and 7 in the figure represent such a pair. On



Titanite discordia. Encyclopædia Britannica, Inc.

this diagram, the presence of surface-correlated lead loss will displace data from the titanite line toward the time of loss close to zero age on the concordia curve. The uranium-lead data would then plot below the line shown, and neither the primary nor secondary age would be defined. The importance of of eliminating recent loss, therefore, is clear. If these ages had been measured by any of the other schemes that have only a single parent-daughter pair, a whole series of different numbers spanning the time from 1.65 billion to 988 million years ago would be observed. There would be no way of telling which of the measured ages, if any, was valid.

Uranium-lead dating relies on the isolation of very high-quality grains or parts of mineral grains that are extremely rare but nevertheless present in most igneous, metamorphic, and sedimentary rock units. Samples weighing 10 to 50 kilograms are collected, crushed, and ground into a fine sand, and the various minerals are isolated on the basis of specific gravity, grain size, and magnetic properties. The minerals used are not visible in the field, but their presence can be inferred from the easily identified major minerals present.

One of the most interesting applications of the improved uranium-lead zircon technique has to do with its ability to achieve nearly concordant results from single grains extracted from sandstone. This is possible because zircon is chemically inert and is not disturbed during weathering and because single grains with a diameter about the thickness of a human hair contain sufficient uranium and lead for analysis in the most advanced laboratories. In one sample it was determined that a sandstone that underlies most of the province of Nova Scotia in Canada was probably originally deposited off the coast of North Africa and thrust over the continent before the opening of the Atlantic Ocean. This follows because the ages observed occur in North Africa, whereas those common in North America are absent.

Another sample, this one from sandstone deposited by a large river in northern Scotland, must have been derived from continental rocks whose ages are represented by those determined for the individually dated sand grains. In this case, the continent from which the sand was derived has moved away as a result of continental drift, but it can be identified by the ages measured.

The Rubidium-Strontium Method

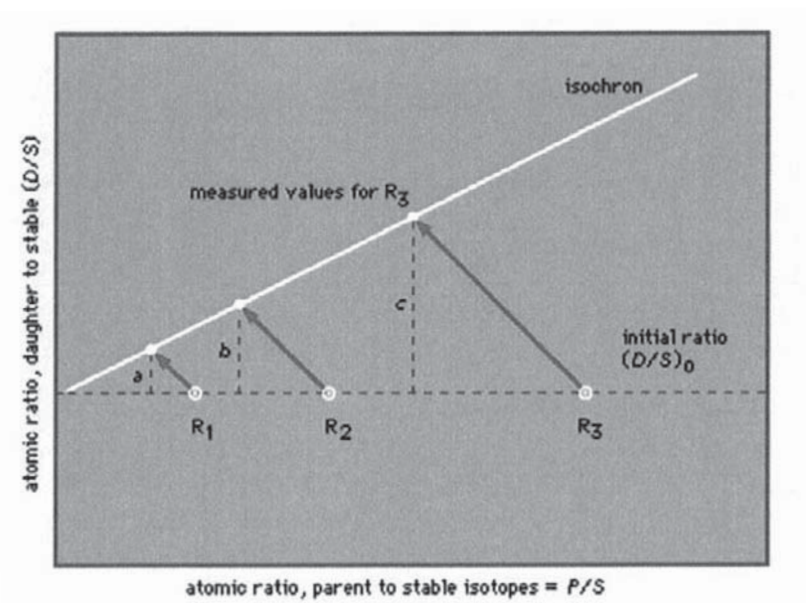
The rubidium-strontium method is used to estimate the age of rocks, minerals, and meteorites from measurements of the amount of the stable isotope strontium-87 formed by the decay of the unstable isotope rubidium-87 that was present in the rock at the time of its formation. Rubidium-87 comprises 27.85 percent of the total atomic abundance of rubidium, and of the four isotopes of strontium, only strontium-87 is formed by its decay. The method is applicable to very old rocks because the transformation is extremely slow. The half-life, or time required for half the initial quantity of rubidium-87 to disappear, is approximately 50 billion years.

Most minerals that contain rubidium also had some strontium incorporated when the mineral was formed, so a correction must be made for this initial amount of strontium to obtain the radiogenic increment (that is, the increase due to decay of rubidium-87).

The radioactive decay of rubidium-87 (^{87}Rb) to strontium-87 (^{87}Sr) was the first widely used dating system that utilized the isochron method. Rubidium is a relatively abundant trace element in the Earth's crust and can be found in many common rock-forming minerals in which it substitutes for the major element potassium. Because rubidium is concentrated in crustal rocks, the continents have a much higher abundance of the daughter isotope strontium-87 compared with the stable isotopes. This relative abundance is expressed as the $^{87}Sr/^{86}Sr$ ratio, where strontium-86 is chosen to represent the stable isotopes strontium-88, strontium-86, and strontium-84, which occur in constant proportions in natural materials. Thus, a precise measurement of the $^{87}Sr/^{86}Sr$ ratio in a modern volcano can be used to determine age if recycled older crust is present. A ratio for average continental crust of about 0.72 has been determined by measuring strontium from clamshells from the major river systems. In contrast, the Earth's most abundant lava rocks, which represent the mantle and make up the major oceanic ridges, have values between 0.703 and 0.705. This difference may appear small, but considering that modern instruments can make the determination to a few parts in 70,000, it is quite significant. Dissolved strontium in the oceans today has a value of 0.709 that is dependent on the relative input from the continents and the ridges. In the geologic past changes in the activity of these two sources has produced varying $^{87}Sr/^{86}Sr$ ratios over time. Thus if well-dated, unaltered fossil shells containing strontium from ancient seawater are analyzed, changes in this ratio with time can be observed and applied in reverse to estimate the time when fossils of unknown age were deposited.

Dating Simple Igneous Rocks

The rubidium-strontium pair is ideally suited for the isochron dating of igneous rocks. As a liquid rock cools, first one mineral and then another achieves saturation and precipitates, each extracting specific elements in the process. Strontium is extracted in many minerals that are formed early, whereas rubidium is gradually concentrated in the final liquid phase. At the time of crystallization, this produces a wide range in the Rb/Sr ratio in rocks that have



Isochron diagram. Encyclopædia Britannica, Inc.

identical $^{87}Sr/^{86}Sr$ ratios. On the isochron diagram shown in the figure on page 128, the samples would plot initially at points R_1 to R_3 along a line representing the initial ratio designated $(^{87}Sr/^{86}Sr)_o$. Over geologic time, this ratio is increased in proportion to the $^{87}Rb/^{86}Sr$ ratio, as discussed earlier, and the line rotates with a slope equal to $(e^{\lambda t} - 1)$ that represents the time elapsed. Thus, the present-day ratio $(^{87}Sr/^{86}Sr)_p$ equals the initial ratio $(^{87}Sr/^{86}Sr)_o$ plus radiogenic additions, or $(^{87}Sr/^{86}Sr)_p = (^{87}Sr/^{86}Sr)_o + {^{87}Rb}/{^{86}Sr}\,(e^{\lambda t} - 1)$. This equation is that of a straight line of the form $y = b + xm$, where $y = (^{87}Sr/^{86}Sr)_p$, the value measured today; b represents $(^{87}Sr/^{86}Sr)_o$, the value initially present; x stands for the $^{87}Rb/^{86}Sr$ ratio; and m is the slope of the line $(e^{\lambda t} - 1)$.

In practice, rock samples weighing several kilograms each are collected from a suite of rocks that are believed to have been part of a single homogeneous liquid prior to solidification. The samples are crushed and homogenized to produce a fine representative rock powder from which a fraction of a gram is withdrawn and dissolved in the presence of appropriate isotopic traces, or spikes. Strontium and rubidium are extracted and loaded into the mass spectrometer, and the values appropriate to the x and y coordinates are calculated from the isotopic ratios measured. Once plotted as $R1_p$ (that is, rock 1 present values), $R2_p$, and $R3_p$, the data are examined to assess how well they fit the required straight line. Using estimates of measurement precision, the crucial question of whether or not scatter outside of measurement error exists is addressed. Such scatter would constitute a geologic component, indicating that one or more of the underlying assumptions has been violated and that the age indicated is probably not valid. For an isochron to be valid, each sample tested must (1) have had the same initial ratio, (2) have been a closed system over geologic time, and (3) have the same age.

Well-preserved, unweathered rocks that crystallized rapidly and have not been subjected to major reheating events are most likely to give valid isochrons. Weathering is a disturbing influence, as is leaching or exchange by hot crustal fluids, since many secondary minerals contain rubidium. Volcanic rocks are most susceptible to such changes because their minerals are fine-grained and unstable glass may be present. On the other hand, meteorites that have spent most of their time in the deep freeze of outer space can provide ideal samples.

Dating Minerals

Potassium-bearing minerals including several varieties of mica, are ideal for rubidium-strontium dating as they have abundant parent rubidium and a low abundance of initial strontium. In most cases, the changes in the $^{87}Sr/^{86}Sr$ ratio are so large that an initial value can be assumed without jeopardizing the accuracy of the results. When minerals with a low-rubidium or a high-strontium content are analyzed, the isochron-diagram approach can be used to provide an evaluation of the data. As discussed above, rubidium-strontium mineral ages need not be identical in a rock with a complex thermal history, so that results may be meaningful in terms of dating the last heating event but not in terms of the actual age of a rock.

Dating Metamorphic Rocks

Should a simple igneous body be subjected to an episode of heating or of deformation or of a combination of both, a well-documented special data pattern develops. With heat, daughter isotopes diffuse out of their host minerals but are incorporated into other minerals in the rock. Eventually the $^{87}Sr/^{86}Sr$ ratio in the minerals becomes identical. When the rock again cools, the minerals close and again accumulate daughter products to record the time since the second event. Remarkably, the isotopes remain within the rock sample analyzed, and so a suite of whole rocks can still provide a valid primary age. This situation is easily visualized on an isochron diagram, where a series of rocks plots on a steep line showing the primary age, but the minerals in each rock plot on a series of parallel lines that indicate the time since the heating event. If cooling is very slow, the minerals with the lowest blocking temperature, such as biotite mica, will fall below the upper end of the line.

A more dramatic presentation of this phenomenon is found when the changes in the $^{87}Sr/^{86}Sr$ ratios in a variety of minerals in a single rock are depicted as a function of geologic time. Here, an essentially rubidium-free, strontium-rich phase like apatite retains its initial $^{87}Sr/^{86}Sr$ ratio over time, whereas the value in such rubidium-rich, strontium-poor minerals as biotite increases rapidly with time. The rock itself gives the integrated, more gradual increase. At the time of heating, identical $^{87}Sr/^{86}Sr$ ratios are again achieved as described above, only to be followed by a second episode of isotopic divergence.

Approaches to this ideal case are commonly observed, but peculiar results are found in situations where the heating is minimal. If one assumes for a moment that only the mineral with the lowest blocking temperature loses its daughter isotope, it is easy to imagine that other low-temperature minerals formed at this time may acquire extremely high $^{87}Sr/^{86}Sr$ ratios. Epidote, a low-temperature alteration mineral with a very high concentration of radiogenic strontium, has been found in rocks wherein biotite has lost strontium by diffusion. The rock itself has a much lower ratio, so that it did not take part in this exchange.

Although rubidium-strontium dating is not as precise as the uranium-lead method, it was the first to be exploited and has provided much of the prevailing knowledge of Earth history. The procedures of sample preparation, chemical separation, and mass spectrometry are relatively easy to carry out, and datable minerals occur in most rocks. Precise ages can be obtained on high-level rocks (that is, those closer to the surface) and meteorites, and imprecise but nevertheless valuable ages can be determined for rocks that have been strongly heated. The mobility of rubidium in deep-level crustal fluids and melts that can infiltrate other rocks during metamorphism as well as in fluids involved in weathering can complicate the results.

The Samarium-Neodymium Method

The radioactive decay of samarium of mass 147 (^{147}Sm) to neodymium of mass 143 (^{143}Nd) has been shown to be capable of providing useful isochron ages for certain geologic materials. Both parent and daughter belong to the rare earth element group, which is itself the subject of numerous geologic investigations. All members of this group have similar chemical properties and charge, but differ significantly in size. Because of this, they are selectively removed as different minerals are precipitated from a melt. In the opposite sense, their relative abundance in a melt can indicate the presence of certain residual minerals during partial melting. Unlike rubidium, which is enriched over strontium in the crust, samarium is relatively enriched with respect to neodymium in the mantle. Consequently, a volcanic rock composed of melted crust would have elevated radiogenic strontium values and depressed radiogenic neodymium values with respect to the mantle. As a parent-daughter pair, samarium-147 and neodymium-143 are unique in that both have very similar chemical properties, and so loss by diffusion may be reduced. Their low concentrations in surface waters indicates that changes during low-temperature alteration and weathering are less likely. Their presence in certain minerals in water-deposited gold veins, however, does suggest mobility under certain conditions. In addition, their behaviour under high-temperature metamorphic conditions is as yet poorly documented.

The exploitation of the samarium-neodymium pair for dating only became possible when several technical difficulties were overcome. Procedures to separate these very similar elements and methods of measuring neodymium isotope ratios with uncertainties of only a few parts in 100,000 had to be developed.

In theory, the samarium-neodymium method is identical to the rubidium-strontium approach. Both use the isochron method to display and evaluate data. In the case of samarium-neodymium dating, however, the chemical similarity of parent and daughter adds another complication because fractionation during crystallization is extremely limited. This makes the isochrons short and adds further to the necessity for high precision. The result is that, with few exceptions, uncertainties in measured ages are as large as 30 to 100 million years, even for the oldest rocks and meteorites. Mineral isochrons provide the best results.

The equation relating present-day neodymium isotopic abundance as the sum of the initial ratios and radiogenic additions is that of a straight line, as discussed earlier for rubidium-strontium.

In a successful application of the samarium-neodymium method to a sample of basalt from the Moon, the constituent minerals plagioclase, ilmenite, and pyroxene provide enough spread in the $^{147}Sm/^{143}Nd$ ratio to allow an age of 3,700 ± 70 million years to be calculated. Other successful examples have been reported where rocks with open rubidium-strontium systems have been shown to have closed samarium-neodymium systems. In other examples, the ages of rocks with insufficient rubidium for dating have been successfully determined. There is considerable promise for dating garnet, a common metamorphic mineral, because it is known to concentrate the parent isotope.

In general, the use of the samarium-neodymium method as a dating tool is limited by the fact that other methods (mainly the uranium-lead approach) are more precise and require fewer analyses. In the case of meteorites and lunar rocks where samples are limited and minerals for other dating methods are not available, the samarium-neodymium method can provide the best ages possible.

The Rhenium-Osmium Method

The decay scheme in which rhenium-187 is transformed to osmium-187 shows promise as a means of studying mantle-crust evolution but has displayed only limited potential for isotopic dating. Technical difficulties have yet to be overcome. Osmium is strongly concentrated in the mantle and extremely depleted in the crust, so that crustal osmium must have exceedingly high radiogenic-to-stable ratios while the mantle values are low. In fact, crustal levels are so low that they are extremely difficult to measure with current technology. Most work to date has centred around rhenium- or osmium-enriched minerals. Another problem that is still under investigation involves the difficulty of ionizing osmium in a mass spectrometer. A number of new approaches are being studied. In one method, success has been demonstrated in ionizing osmium in an instrument called an ion microprobe. This instrument bombards the sample target with a beam of ions of such high energy that atoms in the sample evaporate into a vacuum and a relatively small number becomes ionized. The ions are then extracted into a mass spectrometer, and the desired isotopic abundances are determined. Another related approach being tested involves the evaporation of the sample by a short pulse of ions, with a subsequent pulse of laser light tuned to ionize the desired element. Once the technical difficulties are surmounted, the distribution of rhenium and radiogenic and stable osmium isotopes will be explored. The greatest potential for this method might be in studies concerning the origin and age of mineral deposits since rhenium and osmium are known to occur in such materials.

Potassium-Argon Methods

The radioactive decay scheme involving the breakdown of potassium of mass 40 (^{40}K) to argon gas of mass 40 (^{40}Ar) formed the basis of the first widely used isotopic dating method. Since radiogenic argon-40 was first detected in 1938 by the American geophysicist Lyman T. Aldrich and A.O. Nier, the method has evolved into one of the most versatile and widely employed methods available. Potassium is one of the 10 most abundant elements that together make up 99 percent of the Earth's crust and is therefore a major constituent of many rock-forming minerals. In fact, potassium-40 decays to both argon-40 and calcium-40, but because argon is absent in most minerals while calcium is present, the argon produced is easier to detect and measure. Most of the argon in the Earth's atmosphere has been created by the decay of potassium-40 as the argon-40 abundance is about 1,000 times higher than expected from cosmic abundances. Argon dating involves a different technology from all the other methods so far described because argon exists as a gas at room temperature. Thus it can be purified as it passes down a vacuum line by freezing out or reacting out certain contaminants. It is then introduced into a mass spectrometer through a series of manual or computer-controlled valves. Technical advances, including the introduction of the argon-40–argon-39 method and laser heating, that have improved the versatility of the method, are described below.

In conventional potassium-argon dating, a potassium-bearing sample is split into two fractions: one is analyzed for its potassium content, while the other is fused in a vacuum to release the argon gas. After purification has been completed, a spike enriched in argon-38 is mixed in and the atomic abundance of the daughter product argon-40 is measured relative to the argon-38 added. The amount of the argon-36 present is then determined relative to argon-38 to provide an estimate of the background atmospheric correction. In this case, relatively large samples, which may include significant amounts of alteration, are analyzed. Since potassium is usually added by alteration, the daughter-parent ratio and the age might be too low.

A method designed to avoid such complexities was introduced by the geochronologists Craig M. Merrihue and Grenville Turner in 1966. In this technique, known as the argon-40–argon-39 method, both parent and daughter can be determined in the mass spectrometer as some of the potassium atoms in the sample are first converted to argon-39 in a nuclear reactor. In this way, the problem of measuring the potassium in inhomogeneous samples is eliminated and smaller amounts of material can be analyzed. An additional advantage then becomes possible. The sample can be heated in stages at different temperatures and the age calculated at each step. If alteration is evident, the invalid low-temperature age can be eliminated and a valid high-temperature age determined. In some cases, partly reset systems also may be detected.

As in all dating systems, the ages calculated can be affected by the presence of inherited daughter products. In a few cases, argon ages older than that of the Earth which violate local relative age patterns have even been determined for the mineral biotite. Such situations occur mainly where old rocks have been locally heated, which released argon-40 into pore spaces at the same time that new minerals grew. Under favourable circumstances the isochron method may be helpful, but tests by other techniques may be required. For example, the rubidium-strontium method would give a valid isotopic age of the biotite sample with inherited argon.

As techniques evolved, argon background levels have been reduced and the method has become more and more sensitive. Capitalizing on this, it is now possible to measure the minute amount of argon released when a single spot on a crystal is heated by an intense laser beam. For geologically old potassium-rich materials, a single spot may produce sufficient gas for analysis, whereas single millimetre-sized grains may be required in very young materials. Progressive refinement of the method has made new areas of research possible, and the ability to understand complexities encountered in earlier investigations has increased. In one study the age of volcanic ash as young as 215,000 ± 4,000 years and the presence of inherited older grains in another ash sample were thoroughly documented. This was done by melting single millimetre-sized grains with a laser and measuring individual argon-40–argon-39 ages with a highly sensitive gas mass spectrometer.

The potassium-argon method has provided a great deal of information about the Earth's recent and ancient past. It has been used to measure meteorites as old as 4.5 billion years and volcanic rocks as young as 20,000 years. In addition, the potassium-argon method has been instrumental in determining the ages of the stripes of alternating normally and reversely magnetized volcanic rocks that parallel the axis of the mid-oceanic ridges. In ancient shield areas large segments of crust that were uplifted and cooled at the same time—that is, geologic provinces—have been identified by the potassium-argon method. The technique is highly responsive to thermal events in a relatively predictable fashion, so the cooling history of a region may be established.

Fission-Track Dating

This is a special type of dating method that makes use of a microscope rather than a mass spectrometer, and capitalizes on damaged zones, or tracks, created in crystals during the spontaneous fission of uranium-238. In this unique type of radioactive decay, the nucleus of a single parent uranium atom splits into two fragments of similar mass with such force that a trail of crystal damage is left in the mineral. Immersing the sample in an etching solution of strong acid or base enlarges the fission tracks into tube-shaped holes large enough to be seen under a high-powered microscope. The number of tracks present can be used to calculate the age of the sample if the uranium content is known. Fortunately, the uranium content of precisely the spot under scrutiny can be obtained by a similar process when working with a polished crystal surface. The sample is bombarded with slow (thermal) neutrons in a nuclear reactor, resulting in induced fission of uranium-235 (as opposed to spontaneous fission of uranium-238). The fission tracks produced by this process are recorded by a thin plastic film placed against the surface of the sample. The uranium content of the material can then be calculated so long as the neutron dose is known. The age of the sample is obtained using the equation, age = $N \times Ds/Di \times 6 \times 10^{-8}$, in which N is the total neutron dose expressed as neutrons per square centimetre and Ds is the observed track density for spontaneous fission while Di is that for induced fission.

The preservation of crystal damage (the retention of fission tracks) is highly sensitive to temperature and varies from mineral to mineral. The technique can be used to determine mild thermal events as low as 100° C (212° F). Alternately, primary ages can be calculated if the rock was formed at the surface and cooled quickly. Under these conditions the calculated fission-track ages of two minerals with widely different annealing temperatures would be identical. The accuracy achieved depends on the number of tracks counted, so that artificial glass coloured with 10 percent uranium can be dated as soon as 30 years after manufacture. With uranium levels of a few parts per million, samples as young as 300,000 years can be dated by counting tracks for one hour. When dealing with very old materials, high-uranium samples must be avoided because there are so many interlocking tracks that they can no longer be counted.

A special feature of fission-track dating lies in its ability to map the uranium distribution within mineral grains. In a uranium map for single zircon grains, the outer zones that grew during a major melting event contained much more uranium than the grains originally present. The uranium-lead age was highly biased toward the younger event and the primary age could be determined only after the outer zones were removed. In practice, fission-track dates are regarded as cooling ages unless proved otherwise. It might also be noted that uncertainties in results may arise from an uneven distribution of uranium, statistical errors in counting, and inaccurate estimates of neutron flux (dose of neutrons).

Fission-track dating can be used on a wide variety of minerals found in most geologic materials, and it is relatively inexpensive to apply. Because closure temperatures vary widely from, say, 300° C (572° F) for titanite and zircon to less than 100° C for biotite and apatite, valuable information can be obtained regarding the uplift and cooling rates of crustal rocks.

Carbon-14 Dating and Other Cosmogenic Methods

The occurrence of natural radioactive carbon in the atmosphere provides a unique opportunity to date organic materials as old as 50,000 years. Unlike most isotopic dating methods, the conventional carbon-14 dating technique is not based on counting daughter isotopes. It relies instead on the progressive decay or disappearance of the radioactive parent with time. Carbon-14 has a half-life of 5,730 ± 40 years—that is, half the amount of the radioisotope present at any given time will undergo spontaneous disintegration during the succeeding 5,730 years. Because carbon-14 decays at this constant rate, an estimate of the date at which an organism died can be made by measuring the amount of its residual radiocarbon.

The discovery of natural carbon-14 by Willard Libby of the United States began with his recognition that a process that had produced radiocarbon in the laboratory was also going on in the Earth's upper atmosphere—namely, the bombardment of nitrogen by free neutrons. Newly created carbon-14 atoms were presumed to react with atmospheric oxygen to form carbon dioxide (CO_2) molecules. Radioactive carbon thus was visualized as gaining entrance wherever atmospheric carbon dioxide enters—into land plants by photosynthesis, into animals that feed on the plants, into marine waters and freshwaters as a dissolved component, and from there into aquatic plants and animals. In short, all parts of the carbon cycle were seen to be invaded by the isotope carbon-14.

Invasion is probably not the proper word for a component that Libby calculated should be present only to the extent of about one atom in a trillion stable carbon atoms. So low is such a carbon-14 level that no one had detected natural carbon-14 until Libby, guided by his own predictions, set out specifically to measure it. His success initiated a series of measurements designed to answer two questions: Is the concentration of carbon-14 uniform throughout the plant and animal kingdoms? And, if so, has today's uniform level prevailed throughout the recent past?

After showing the essential uniformity of carbon-14 in living material, Libby sought to answer the second question by measuring the radiocarbon level in organic samples dated historically—materials as old as 5,000 years from sources such as Egyptian tombs. With correction for radioactive decay during the intervening years, such old samples hopefully would show the same starting carbon-14 level as exists today. This was just what Libby's measurements indicated. His conclusion was that over the past 5,000 years the carbon-14 level in living materials has remained constant within the 5 percent precision of measurement. A dating method was thus available, subject only to confirmation by actual application to specific chronologic problems.

$$t = \frac{t_{1/2}}{0.693} \times \log_e\left(\frac{N_0}{N}\right).$$

Since Libby's foundational studies, tens of thousands of carbon-14 measurements of natural materials have been made. Expressed as a fraction of the contemporary level, they have been mathematically converted to ages using the equation above. Archaeology has been the chief beneficiary of radioactive-carbon dating, but late glacial and postglacial chronological studies in geology have also been aided greatly.

Improvements in measurement accuracy and the ever-mounting experience in applying carbon-14 dating have provided superior and more voluminous data with which to better answer Libby's original questions. It is now clear that carbon-14 is not homogeneously distributed among today's plants and animals. The occasional exceptions all involve nonatmospheric contributions of carbon-14-depleted carbon dioxide to organic synthesis. Specifically, volcanic carbon dioxide is known to depress the carbon-14 level of nearby vegetation and dissolved limestone carbonate occasionally has a similar effect on freshwater mollusks, as does upwelling of deep ocean water on marine mollusks. In every case, the living material affected gives the appearance of built-in age.

In addition to spatial variations of the carbon-14 level, the question of temporal variation has received much study. A 2 to 3 percent depression of the atmospheric radioactive-carbon level since 1900 was noted soon after Libby's pioneering work, almost certainly the result of the dumping of huge volumes of carbon-14-free carbon dioxide into the air through smokestacks. Of more recent date was the overcompensating effect of man-made carbon-14 injected into the atmosphere during nuclear-bomb testing. The result was a rise in the atmospheric carbon-14 level by more than 50 percent. Fortunately, neither effect has been significant in the case of older samples submitted for carbon-14 dating. The ultimate cause of carbon-14 variations with time is generally attributed to temporal fluctuations in the cosmic rays that bombard the upper atmosphere and create terrestrial carbon-14. Whenever the number of cosmic rays in the atmosphere is low, the rate of carbon-14 production is correspondingly low, resulting in a decrease of the radioisotope in the carbon-exchange reservoir described above. Studies have revealed that the atmospheric radiocarbon level prior to 1000 BCE deviates measurably from the contemporary level. In the year 6200 BCE, it was about 8 percent above what it is today. In the context of carbon-14 dating, this departure from the present-day level means that samples with a true age of 8,200 years would be dated by radiocarbon as 7,500 years old.

The problems stemming from temporal variations can be overcome to a large degree by the use of calibration curves in which the carbon-14 content of the sample being dated is plotted against that of objects of known age. In this way, the deviations can be compensated for and the carbon-14 age of the sample converted to a much more precise date. Calibration curves have been constructed using dendrochronological data (tree-ring measurements of bristlecone pines as old as 8,200 years); periglacial varve, or lake sediment, data; and, in archaeological research, certain materials of historically established ages. It is clear that carbon-14 dates lack the accuracy that traditional historians would like to have. There may come a time when all radiocarbon ages rest on firmer knowledge of the sample's original carbon-14 level than is now available. Until then, the inherent error from this uncertainty must be recognized.

A final problem of importance in carbon-14 dating is the matter of sample contamination. If a sample of buried wood is impregnated with modern rootlets or a piece of porous bone has recent calcium carbonate precipitated in its pores, failure to remove the contamination will result in a carbon-14 age between that of the sample and that of its contaminant. Consequently, numerous techniques for contaminant removal have been developed. Among them are the removal of humic acids from charcoal and the isolation of cellulose from wood and collagen from bone. Today, contamination as a source of error in samples younger than 25,000 years is relatively rare. Beyond that age, however, the fraction of contaminant needed to have measurable effect is quite small, and, therefore, undetected or unremoved contamination may occasionally be of significance.

A major breakthrough in carbon-14 dating occurred with the introduction of the accelerator mass spectrometer. This instrument is highly sensitive and allows precise ages on as little as one milligram of carbon, where the older method might require as much as 25 grams for ancient material. The increased sensitivity results from the fact that all of the carbon atoms of mass 14 can be counted in a mass spectrometer. By contrast, if carbon-14 is to be measured by its radioactivity, only those few atoms decaying during the measurement period are recorded. By using the accelerator mass spectrometer, possible interference from nitrogen-14 is avoided since it does not form negative ion beams, and interfering molecules are destroyed by stripping electrons away by operating at several million volts.

The development of the accelerator mass spectrometer has provided new opportunities to explore other rare isotopes produced by the bombardment of the Earth and meteorites by high-energy cosmic rays. Many of these isotopes have short half-lives and hence can be used to date events that happened in the past few thousand to a few million years. In one case, the time of exposure, like the removal of rock by a landslide, can be dated by the presence of the rare beryllium-10 (^{10}Be) isotope formed in the newly exposed surface of a terrestrial object or meteoroidal fragment by cosmic-ray bombardment. Other applications include dating groundwater with chlorine-36 (^{36}Cl), dating marine sediments with beryllium-11 (^{11}Be) and aluminum-26 (^{26}Al), and dating glacial ice with krypton-81 (^{81}Kr). In general, the application of such techniques is limited by the enormous cost of the equipment required.

Uranium-Series Disequilibrium Dating

The isotopic dating methods discussed so far are all based on long-lived radioactive isotopes that have survived since the elements were created or on short-lived isotopes that were recently produced by cosmic-ray bombardment. The long-lived isotopes are difficult to use on young rocks because the extremely small amounts of daughter isotopes present are difficult to measure. A third source of radioactive isotopes is provided by the uranium- and thorium-decay chains. The decay of uranium to lead is not achieved by a single step but rather involves a whole series of different elements, each with its own unique set of chemical properties.

In closed-system natural materials, all of these intermediate daughter elements exist in equilibrium amounts. That is to say, the amount of each such element present is constant and the number that form per unit time is identical to the number that decay per unit time. Accordingly, those with long half-lives are more abundant than those with short half-lives. Once a uranium-bearing mineral breaks down and dissolves, the elements present may behave differently and equilibrium is disrupted. For example, an isotope of thorium is normally in equilibrium with uranium-234 but is found to be virtually absent in modern corals even though uranium-234 is present. Over a long period of time uranium-234, however, decays to thorium-230, which results in a build-up of the latter in old corals and thereby provides a precise measure of time.

Most of the studies using the intermediate daughter elements were for years carried out by means of radioactive counting techniques—that is, the number of atoms present was estimated by the radioactivity of the sample. The introduction of highly sensitive mass spectrometers that allow the total number of atoms to be measured rather than the much smaller number that decay has resulted in a revolutionary change in the family of methods based on uranium and thorium disequilibrium.

Thorium-230 Dating

The insoluble nature of thorium provides for an additional disequilibrium situation that allows sedimentation rates in the modern oceans to be determined. In this case, thorium-230 in seawater, produced principally by the decay of uranium-234, is deposited preferentially in the sediment without the uranium-234 parent. This is defined as excess thorium-230 because its abundance exceeds the equilibrium amount that should be present. With time, the excess decays away and the age of any horizon in a core sample can be estimated from the observed thorium-230-to-thorium-232 ratio in the seawater-derived component of the core. Sedimentation rates between 1 and 20 millimetres (0.04 and 0.8 inch) per 1,000 years are commonly found with slight variations between the major ocean basins.

Lead-210 Dating

The presence of radon gas as a member of the uranium-decay scheme provides a unique method for creating disequilibrium. The gas radon-222 (^{222}Rn) escapes from the ground and decays rapidly in the atmosphere to lead-210 (^{210}Pb), which falls quickly to the surface where it is incorporated in glacial ice and sedimentary materials. By assuming that the present deposition rate also prevailed in the past, the age of a given sample at depth can be estimated by the residual amount of lead-210. Lead-210 dating is particularly useful for determining the ages of relatively recent lacustrine and coastal marine sediments. As a result, it has been applied increasingly to studies concerned with the impact of human activity on the aquatic environment (such as measuring the accumulation rates of pollutants in sediments).

OTHER DATING METHODS AND TECHNIQUES

Several additional dating methods have been used for specific purposes. Most of the techniques listed below involve the decay of certain chemical elements. However, dendrochronology involves the measurement of growth rings in plants.

Dendrochronology

Dendrochronology, which is also called tree-ring dating, is the scientific discipline concerned with dating and interpreting past events, particularly paleoclimates and climatic trends, based on the analysis of tree rings. Samples are obtained by means of an increment borer, a simple metal tube of small diameter that can be driven into a tree to get a core extending from bark to centre. This core is split in the laboratory, the rings are counted and measured, and the sequence of rings is correlated with sequences from other cores.

Dendrochronology is based on the fact that many species of trees produce growth rings during annual growing seasons. The width of the ring (the amount of growth) for each year is determined by various internal and external factors, but it tends to vary mainly in proportion to

These bristlecone pines in the Inyo National Forest near Bishop, California, are 4,700 years old. They are the oldest trees still living in the world. Gabriel Bouys/AFP/Getty Images either the amount of available precipitation or the prevailing temperatures. The ring measurements taken from trees with overlapping ages can extend knowledge of climates back thousands of years. The bristlecone pines of California have proven to be particularly suitable for such chronologies, since some individual trees are more than 4,000 years old.

Helium Dating

Helium dating is a method of age determination that depends on the production of helium during the decay of the radioactive isotopes uranium-235, uranium-238, and thorium-232. Because of this decay, the helium content of any mineral or rock capable of retaining helium will increase during the lifetime of that mineral or rock, and the ratio of helium to its radioactive progenitors then becomes a measure of geologic time. If the parent isotopes are measured, the helium dating method is referred to as uranium-thorium-helium dating. If only the alpha-particle emission and helium content are measured, the method is called the alpha-helium radioactive clock. Alpha particles are the nuclei of helium atoms emitted from the nucleus of the radioactive progenitor.

Before the use of mass spectrometry in isotopic geochronology, helium dating provided most of the dates used in the early geologic time scales. Helium ages, however, tend to be too low because the gas escapes from the rock. A thermal event that will leave most radioactive clocks relatively unaffected may have a drastic effect on the helium radioactive clock. In the future, helium dating may be found very useful for dating rocks of the late Cenozoic and Pleistocene, because rocks and minerals of this age have not been subject to the complex history of older rocks and minerals. Thus, all the helium is more likely to have been retained. Fossils, as well as minerals and rocks, may be dated by helium dating. The relatively large amount of helium produced in rocks may make it possible to extend helium dating to rocks and minerals as young as a few tens of thousands of years old.

Ionium-Thorium Dating

The method of establishing the time of origin of marine sediments according to the amount of ionium and thorium they contain is called ionium-thorium dating.

Because uranium compounds are soluble in seawater, while thorium compounds are quite insoluble, the thorium isotopes produced by the decay of uranium in seawater are readily precipitated and incorporated in sediments. One of these thorium isotopes, thorium-230 (also known as ionium), has a half-life of about 80,000 years, which makes it suitable for dating sediments as old as 400,000 years. Thus, the amount of ionium in sediments can be used as a rough measure of the age of sediment. Accurate dating by measurement of ionium alone requires that the rate of sedimentation of ionium be constant with time, an assumption that does not hold for many sediments. Any thorium-232 present in seawater will also precipitate, and the decay of the ratio of ionium to thorium-232 can be used as a measure of time. This method does not require a constant rate of sedimentation of ionium but simply that the two isotopes are precipitated in a constant proportion.

The validity of the ionium-thorium age method is based upon the following assumptions: (1) the ratio of thorium-230 (ionium) to thorium-232 in ocean water remains constant during the age span of the sample to be dated; (2) during precipitation there is no chemical fractionation of ionium and thorium, so that they are precipitated in a constant ratio; (3) the sediments do not contain any detrital material that has a significant amount of either nuclide; and (4) after deposition the thorium isotopes do not migrate within the sediments. When these assumptions are valid, an accurate date may be obtained for the age of marine sediments.

Obsidian-Hydration-Rind Dating

This is a method of age determination for obsidian (black volcanic glass) that makes use of the fact that obsidian freshly exposed to the atmosphere will take up water to

Obsidian, shown here, is a glassy volcanic, or igneous, rock. Wally Eberhart/Visuals Unlimited/Getty Images form a hydrated surface layer with a density and refractive index different from that of the remainder of the obsidian. The thickness of the layer can be determined by microscopic examination of a thin section of the sample cut at right angles to the surface. The hydration-rind dating technique also has been used to date glassy rhyolitic flows that have erupted more than 200 years ago but less than 200,000 years ago.

Protactinium-231–Thorium-230 Dating

This method of age determination makes use of the quantities of certain protactinium and thorium isotopes in a marine sediment. Protactinium and thorium have very similar chemical properties and appear to be precipitated at the same rates in marine sediments. The isotopes protactinium-231 and thorium-230 are both radioactive and decay with half-lives of 32,500 years and 80,000 years, respectively. The ratio of the two radioactive isotopes constitutes a better radioactive geochronometer than either of them separately, because they do not need to have a uniform sedimentation rate through time but need only be precipitated in the same proportion. It is likely that this condition will hold even though the rate of sedimentation may vary. Sediments as old as 175,000 years may be dated by this method.

Radiation-Damage Dating

Radiation-damage dating is a method of age determination that makes use of the damage to crystals and the radiation from radioactive substances caused by storage of energy in electron traps. In the mineral zircon, for example, radiation damage results in a change in colour, the storage of energy in electron traps, and a change in the crystallographic constants of the mineral. Extensive damage may result in a metamict mineral (that is, a mineral in which the crystal structure has been destroyed). The change in crystallographic constants is a function of the total radiation damage, which depends on the amount of radioactive substances and the age of the mineral. Thus, measurement of uranium and thorium content in the zircon, combined with measurement of its crystallographic constants, provides a measure of its age.

Uranium-234–Uranium-238 Dating

This method of age determination makes use of the radioactive decay of uranium-238 to uranium-234. The method can be used for dating of sediments from either a marine or a playa lake environment. Because this method is useful for the period of time from about 100,000 years to 1,200,000 years before the present, it helps in bridging the gap between the carbon-14 dating method and the potassium-argon dating method.

Uranium-Thorium-Lead Dating

This method, which is also called common-lead dating, establishes the time of origin of a rock by means of the amount of common lead it contains. Common lead is any lead from a rock or mineral that contains a large amount of lead and a small amount of the radioactive progenitors of lead—specifically, the uranium isotopes uranium-235 and uranium-238 and the thorium isotope thorium-232.

The important characteristic of common lead is that it contains no significant proportion of radiogenic lead

This piece of anorthrosite was found in the high peaks of the Adirondack Mountains of New York. Ted Kinsman/Photo Researchers, Inc.

accumulated since the time that the mineral or rock phase was formed. Of the four isotopes of lead, two are formed from the uranium isotopes and one is formed from the thorium isotope. Only lead-204 is not known to have any long-lived radioactive progenitor. Primordial lead is thought to have been formed by stellar nuclear reactions, released to space by supernovae explosions, and incorporated within the dust cloud that constituted the primordial solar system. The troilite (iron sulfide) phase of iron meteorites contains lead that approximates the primordial composition. The lead incorporated within the Earth has been evolving continuously from primordial lead and from the radioactive decay of uranium and thorium isotopes. Thus, the lead isotopic composition of any mineral or rock depends upon its age and the environment from which it was formed. That is, it would depend upon the ratio of uranium plus thorium to lead in the parent material.

The Earth can be assumed to be a very large sample containing lead evolving from primordial lead by radiogenic increments. If modern lead, for example, from marine sediments or modern basalts has the composition of lead in the Earth and if the lead in the troilite phase of iron meteorites has the composition of primordial lead, then a simple model yields about 4.6 billion years for the age of the Earth. This age is in good agreement with the age of the meteorites and the age of the Moon as determined independently.

CHAPTER 3

Precambrian Time

Precambrian time is the period of time that extends from about 4 billion years ago (the approximate age of the oldest known rocks) to the beginning of the Cambrian Period, 542 million years ago. The Precambrian represents more than 80 percent of the total geologic record.

All life-forms were long assumed to have originated in the Cambrian, and therefore all earlier rocks were grouped together into the Precambrian. Although many varied forms of life evolved and were preserved extensively as fossil remains in Cambrian sedimentary rocks, detailed mapping and examination of Precambrian rocks on most continents have revealed that additional primitive life-forms existed more than 3.5 billion years ago. Nevertheless, the original terminology to distinguish Precambrian rocks from all younger rocks is still used for subdividing geologic time.

The earliest evidence for the advent of life includes Precambrian microfossils that resemble algae, cysts of flagellates, tubes interpreted to be the remains of filamentous organisms, and stromatolites (sheetlike mats precipitated by communities of microorganisms). In the late Precambrian, the first multicellular organisms evolved, and sexual division developed. By the end of the Precambrian, conditions were set for the explosion of life that took place at the start of the Phanerozoic Eon.

EONS OF PRECAMBRIAN TIME

The Precambrian is divided into three eons. However, only two (the Archean and Proterozoic) have been formally

Stromatolites, such as those shown here, are the fossil remains of early Proterozoic bacteria. Pete Ryan/National Geographic/Getty Images

delineated. The third eon, the Hadean, is an unofficial division of time.

The Hadean Eon

The Hadean Eon is an informal division of Precambrian time occurring between about 4.6 billion and about 4.0 billion years ago. The Hadean Eon is characterized by Earth's initial formation—from the accretion of dust and gases and the frequent collisions of larger planetesimals—and by the stabilization of its core and crust and the development of its atmosphere and oceans. Throughout part of the eon, impacts from extraterrestrial bodies released enormous amounts of heat that likely prevented much of the rock from solidifying at the surface. As such, the name of the interval is a reference to Hades, a Greek translation of the Hebrew word for hell.

Earth's surface was incredibly unstable during the early part of the Hadean Eon. Convection currents in the mantle brought molten rock to the surface and caused cooling rock to descend into magmatic seas. Heavier elements, such as iron, descended to become the core, whereas lighter elements, such as silicon, rose and became incorporated into the growing crust. Although no one knows when the first outer crust of the planet formed, some scientists believe that the existence of a few grains of zircon dated to about 4.4 billion years ago confirm the presence of stable continents, liquid water, and surface temperatures that were probably less than 100 °C (212 °F). Since Hadean times, nearly all of this original crust has subducted from the movements of tectonic plates, and thus few rocks and minerals remain from the interval. The oldest rocks known are the faux amphibolite volcanic deposits of the Nuvvuagittuq greenstone belt in Quebec, Canada. They are estimated to be 4.28 billion years old. The oldest minerals are the aforementioned grains of zircon, which were found in the Jack Hills of Australia.

Considerable debate surrounds the timing of the formation of the atmosphere, as well as its initial composition. Although many scientists contend that the atmosphere and the oceans formed during the latter part of the eon, the discovery of the zircon grains in Australia provide compelling evidence that the atmosphere and ocean formed before 4.4 billion years ago. The early atmosphere likely began as a region of escaping hydrogen and helium. It is generally thought that ammonia, methane, and neon were present sometime after the crust cooled, and volcanic outgassing added water vapour, nitrogen, and additional hydrogen. Some scientists state that ice delivered by comet impacts could have supplied the planet with additional water vapour. Later, it is thought, much of the water vapour in the atmosphere condensed to form clouds and rain that left large deposits of liquid water on Earth's surface.

The Moon is also thought to have formed during the Hadean Eon, and several theories of the Moon's origin have been posited. The leading theory asserts that a collision between Earth and a celestial body the size of Mars ejected material that eventually coalesced into the Moon.

The Archean Eon

The Archean Eon is the earlier of the two official divisions of Precambrian time, occurring between about 4.0 billion and 2.5 billion years ago. Some scientists contend that the Archean Eon began with the formation of the Earth's crust. Records of Earth's primitive atmosphere and oceans emerge in the earliest Archean (Eoarchean Era), and evidence of the earliest primitive life-forms—bacteria and blue-green algae—appears in rocks about 3.5 billion years old. Archean greenstone-granite belts contain many economic mineral deposits, including gold and silver.

It is thought that the oxygen content in today's atmosphere must have slowly accumulated through time starting with an atmosphere that was anoxic during Archean times. Although volcanoes exhale much water vapour (H_2O) and carbon dioxide (CO_2), the amount of free oxygen (O_2) emitted is very small. The inorganic breakdown (photodissociation) of volcanic-derived water vapour and carbon dioxide in the atmosphere would have produced only a small amount of free oxygen. The bulk of the free oxygen in the Archean atmosphere was derived from organic photosynthesis of carbon dioxide (CO_2) and water (H_2O) by anaerobic cyanobacteria (blue-green algae), a process that releases oxygen as a by-product. These organisms were prokaryotes, a group of unicellular organisms with rudimentary internal organization.

Archean oceans were likely created by the condensation of water derived from the outgassing of abundant volcanoes. Iron was released then (as today) into the oceans from submarine volcanoes in oceanic ridges and during the creation of thick oceanic plateaus. This ferrous iron (Fe^{2+}) combined with oxygen and was precipitated as ferric iron in hematite (Fe_2O_3), which produced banded-iron formations on the flanks of the volcanoes. The transfer of biologically produced oxygen from the atmosphere to the sediments was beneficial to the photosynthetic organisms, because at the time free oxygen was toxic to them. When banded-iron formations were being deposited, oxygen-mediating enzymes had not yet developed. Therefore, this removal of oxygen allowed early anaerobes (life-forms not requiring oxygen for respiration) to develop in the early oceans of the Earth.

Carbon dioxide emissions are abundant from modern volcanoes, and it is assumed that the intense volcanism during the Archean Eon caused this gas to be highly concentrated in the atmosphere. This high concentration most likely gave rise to an atmospheric greenhouse effect that warmed the Earth's surface sufficiently to prevent the development of glaciations, for which there is no evidence in Archean rocks. The CO_2 content in the atmosphere has decreased over geological time, because much of the oxygen formerly bound in CO_2 has been released to provide increasing amounts of O_2 to the atmosphere. In contrast, carbon has been removed from the atmosphere via the burial of organic sediments.

Throughout the Archean, oceanic and island arc crust was produced semi-continuously for 1.5 billion years. Thus, most Archean rocks are igneous. The oldest known rocks on Earth, estimated at 4.28 billion years old, are the faux amphibolite volcanic deposits of the Nuvvuagittuq greenstone belt in Quebec, Canada. The second oldest rocks are the 4-billion-year-old Acasta granitic gneisses in northwestern Canada, and a single relict zircon grain dated to 4.2 billion years ago was found within these gneisses. Other ancient sediments and lavas occur in the 3.85-billion-year-old Isua belt of western Greenland (which is similar to an accretionary wedge in the trench of a modern subduction zone) and the 3.5-billion-year-old Barberton Complex in South Africa, which is probably a slice of oceanic crust. A huge pulse in the formation of island arcs and oceanic plateaus took place worldwide from 2.9 to 2.7 billion years ago.

Archean rocks mostly occur in large blocks hundreds to thousands of kilometres across, such as in the Superior and Slave provinces in Canada; the Pilbara and Yilgarn blocks in Australia; the Kaapvaal craton in southern Africa; the Dharwar craton in India; the Baltic, Anabar, and Aldan shields in Russia; and the North China craton. Smaller relicts of Archean rocks in various stages of obliteration occur in many younger Proterozoic and Phanerozoic orogenic (mountain) belts. Some Archean rocks that occur in greenstone-granite belts (zones rich in volcanic rocks that are primitive types of oceanic crust and island arcs) formed on or near the surface of the Earth and thus preserve evidence of the early atmosphere, oceans, and life-forms. Other rocks that occur in granulite-gneiss belts (zones of rocks that were metamorphosed in the Archean mid-lower crust) are exhumed remnants of the lower parts of the Archean continents and thus preserve evidence of deep crustal processes operating at the time.

In greenstone-granite belts there are many oceanic lavas, island arcs, and oceanic plateaus. Therefore, they commonly contain rock types such as basalts, andesites, rhyolites, granitic plutons, oceanic cherts, and ultramafic komatiites (lavas enriched in magnesium, a special product of the melting of the hot Archean mantle). These igneous rocks are host to multitudes of economic mineral deposits of gold, silver, chromium, nickel, copper, and zinc, which are the mainstay of the economies of Canada, Australia, and Zimbabwe.

In granulite-gneiss belts the roots of many Andean-type active continental margins are exposed, the rocks being highly deformed and recrystallized during metamorphism in the deep crust. Common rocks are tonalites (a granitic-type rock rich in plagioclase feldspar) transformed into tonalitic gneisses, amphibolite dikes, and amphibolites derived from volcanic activity. Few mineral deposits occur in granulite-gneiss belts, in common with the deep crust of younger orogenic belts, which are relatively barren of ore concentrations.

The Proterozoic Eon

The Proterozoic Eon is the younger of the two official divisions of Precambrian time. It extended from 2.5 billion to 542 million years ago and is often divided into the Paleoproterozoic (2.5 billion to 1.6 billion years ago), the Mesoproterozoic (1.6 billion to 1 billion years ago), and the Neoproterozoic (1 billion to 542 million years ago) eras. Proterozoic rocks have been identified on all the continents and often constitute important sources of metallic ores, notably of iron, gold, copper, uranium, and nickel. During the Proterozoic, the atmosphere and oceans changed significantly. Proterozoic rocks contain many definite traces of primitive life-forms—the fossil remains of bacteria and blue-green algae, as well as of the first oxygen-dependent animals, the Ediacara fauna.

Oxygen is a by-product of photosynthesis. Free oxygen in the atmosphere increased significantly as a result of biological activity during the Proterozoic. The most important period of change occurred between 2.3 billion and 1.8 billion years ago when free oxygen began to accumulate in the atmosphere. Oxygen levels fluctuated during this time, coinciding with the peak deposition period of banded-iron formations, which removed surplus oxygen from the atmosphere throughout the world. Ferrous iron (Fe^{2+}) in the oceans combined with atmospheric oxygen and, after oxidizing to Fe_2O_3, precipitated as the mineral hematite on the ocean floor. Continued biological activity allowed atmospheric oxygen concentrations to increase.

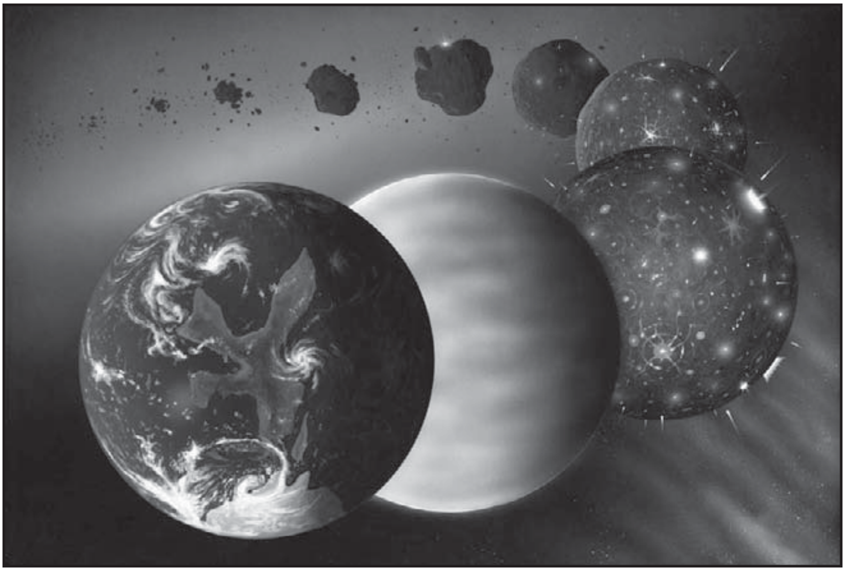
By the time eukaryotes became established in the environment, atmospheric oxygen pressure had risen from low values to about 10 percent of the present atmospheric level (PAL). Megascopic eukaryotes first appeared about 2.3 billion years ago and became widespread by about 1.8 billion years ago. Eukaryotes employed a form of respiration and oxidative metabolism. They had a central nucleus that could split into separate sex cells, and so for the first time a mixed and variable genetic code could be passed to younger generations.

Early organisms on Earth flourished most easily in the shallow water of continental margins. Such stable continental shelf environments, which were rare in the Archean, developed after 2.5 billion years ago, facilitating the growth of photosynthetic organisms and thus oxygen production. Evidence of the rapid rise in the oxygen content includes the first appearance on continental margins of red sandstones. Their colour is caused by the coating of quartz grains with hematite. Other evidence is provided by the occurrence of hematite-rich fossil soil beds that date to about 2.5 billion years ago. The formation of these beds is consistent with a drastic rise in oxygen pressure to 0.1 atmosphere (100 millibars) between 2.2 billion and 2.0 billion years ago.

By 600 million to 543 million years ago, the multicellular Ediacara fauna had appeared. These were the first metazoans (animals made up of more than one type of cell) that required oxygen for growth. The soft-bodied Ediacara fauna were the precursors of organisms with skeletons, the appearance of which marked the end of the Proterozoic and the beginning of the Phanerozoic Eon.

The history of the Proterozoic Eon is dominated by the formation and breakup of supercontinents. By the time of the Archean-Proterozoic boundary about 2.5 billion years ago, many small cratons (stable interior portions of continents) dominated by island arcs had coalesced into one large landmass, or supercontinent. The breakup of this landmass is indicated by the intrusion of abundant transcontinental swarms of dolerite (a type of fine-grained igneous rock) dikes during the period of 2.4 billion to 2.2 billion years ago. These dikes resulted from the impingement of mantle plumes onto the base of the continental crust. This was the fundamental cause of the breakup of the initial supercontinent. During the period between 2.1 billion and 1.8 billion years ago, these fragments again coalesced, by collision tectonics, into a new supercontinent called Columbia. Modern plate-tectonic processes were in operation by at least 2.1 billion to 2.0 billion years ago, as shown by two of the world's oldest well-preserved ophiolites (fragments of oceanic crust), located in the Purtuniq complex in Labrador and the Jourma complex in Finland. The fragmentation of Columbia gave rise to many smaller continents that had eventually assembled into another supercontinent, or a group of several large continental pieces in close proximity to one another, by about 1.0 billion years ago. This assemblage is called Rodinia.

Rodinia was intruded by many basaltic dikes after 1.0 billion years ago. These dikes contributed to the supercontinent's fragmentation and were associated with the



This illustration shows stages in the formation of Earth. The globe in front shows the ancient continent of Rodinia during the Precambrian time. Richard Bizley/Photo Researchers, Inc.

formation of the Iapetus Ocean about 600 million years ago. Other indications of plume activity and continental breakup are vast piles of basalts and transcontinental rifts. A key example is the 1.1-billion-year-old Keweenawan Rift in North America that extends from Michigan via Lake Superior to Kansas. This rift, which is 2,000 km (about 1,200 miles) long and 160 km (100 miles) wide, contains a pile of basaltic lavas 25 km (about 16 miles) thick.

Many mountain belts formed during the Proterozoic, in particular during the intervals between 2.1 and 1.8 billion, 1.3 and 1.0 billion, and 800 and 500 million years ago, associated with the breakup of supercontinents and the subsequent collision of their fragments. New ocean basins were created by the rifting apart of the continents and were subsequently destroyed in subduction zones similar to those under modern-day Japan. The closure of these oceans enabled continental blocks to collide, giving rise to major mountain belts such as the Grenville belt in eastern North America. This belt, which is 1.3 to 1.0 billion years old and 4,000 km (about 2,500 miles) long, was very similar in origin to the Himalayan Mountains that formed in recent geological time. Other major Proterozoic mountain belts created by continental collisions include the Wopmay Orogen in northwest Canada (2.1 billion years old), the Trans-Hudson in Canada (1.8 billion years old), the Svecofennian in Finland (1.9 to 1.8 billion years old), the Ketilidian Orogen (1.8 billion years old) in southwestern Greenland, and the Braziliano, Namibian, and Mozambique belts, which are all about 900 to 500 million years old. In contrast, mountain belts such as the 2.1-billion-year-old Birimian in West Africa and the 1-billion-year-old to 500-million-year-old belts of the Arabian-Nubian Shield developed by the addition of new material largely derived from the Earth's mantle. Thus, they include many island arcs similar to those found in modern-day Japan as well as many ophiolite sequences.

Many Phanerozoic basins contain thick piles of sediments and lie partly to completely on top of Proterozoic mountain belts, obscuring the underlying geological relationships. Some Phanerozoic mountain belts, such as the Himalayas, contain blocks of Proterozoic rocks many tens of kilometres in size that have been heavily reworked by later tectonic activity.

THE PRECAMBRIAN ENVIRONMENT

Several rock types yield information on the range of environments that may have existed during Precambrian time. Evolution of the atmosphere is recorded by banded-iron formations (BIFs), paleosols (buried soil horizons), and red beds, whereas tillites (sedimentary rocks formed by the lithification of glacial till) provide clues to the climatic patterns that occurred during Precambrian glaciations.

Paleogeography

One of the most important factors controlling the nature of sediments deposited today is continental drift. This follows from the fact that the continents are distributed at different latitudes, and latitudinal position affects the temperature of oceanic waters along continental margins (the combined area of the continental shelf and continental slope). In short, sedimentary deposition is climatically sensitive. At present, most carbonates and oxidized red soils are being deposited within 30 degrees of the Equator, phosphorites within 45 degrees, and evaporites within 50 degrees. Most fossil carbonates, evaporites, phosphorites, and red beds of Phanerozoic age dating back to the Cambrian have a similar bimodal distribution with respect to their paleoequators. If the uniformitarian principle that the present is the key to the past is valid (meaning the same geologic processes occurring today occurred in the past), then sediments laid down during the Precambrian would have likewise been controlled by the movement and geographic position of the continents. Thus, it can be inferred that the extensive evaporites dating to 3.5 billion years ago from the Pilbara region of Western Australia could not have been formed within or near the poles. It can also be inferred that stromatolite-bearing dolomites of Riphean rock, a sedimentary sequence spanning the period from 1.65 billion to 800 million years ago, were deposited in warm, tropical waters. Riphean rock is primarily located in the East European craton, which extends from Denmark to the Ural Mountains, and in the Siberian craton in Russia.

Today, phosphate sediments are deposited primarily along the western side of continents. This is the result of high biological productivity in nearby surface waters due to the upwelling of nutrient-rich currents that are moving toward the Equator. The major phosphorite deposits of the Aravalli mountain belt of Rajasthan in northwestern India, which date from the Proterozoic Eon, are associated with stromatolite-rich dolomites. They were most likely deposited on the western side of a continental landmass that resided in the tropics.

Paleoclimate

Earth's climate mechanisms began and evolved during Precambrian times. The atmosphere was a reducing one until photosynthetic organisms evolved to give off oxygen as a byproduct of their metabolism. In the oceans, oxygen produced by cyanobacteria reacted with ferrous iron to form BIFs before it could be released to accumulate in the atmosphere. Like today, climatic conditions of the Precambrian largely depended on continental arrangement and volcanic activity. Large numbers of active volcanoes, such as those that occur when continents break apart, point toward warmer temperatures. Reduced volcanic activity, such as that which occurs during times of continental coalescence and stability, is thought to have stimulated the development of glaciers and ice sheets. Geologic evidence suggests that some glaciations were extensive enough to reach tropical latitudes.

The Evolution of the Atmosphere and Ocean

During the long course of Precambrian time, the climatic conditions of the Earth changed considerably. Evidence of this can be seen in the sedimentary record, which documents appreciable changes in the composition of the atmosphere and oceans over time.

The Oxygenation of the Atmosphere

Earth almost certainly possessed a reducing atmosphere before 2.5 billion years ago. The Sun's radiation produced organic compounds from reducing gases—methane (CH_4) and ammonia (NH_3). The minerals uraninite (UO_2) and pyrite (FeS_2) are easily destroyed in an oxidizing atmosphere. Confirmation of a reducing atmosphere is provided by unoxidized grains of these minerals in 3.0-billion-year-old sediments. However, the presence of many types of filamentous microfossils dated to 3.45 billion years ago in the cherts of the Pilbara region suggests that photosynthesis had begun to release oxygen into the atmosphere by that time. The presence of fossil molecules in the cell walls of 2.5-billion-year-old blue-green algae (cyanobacteria) establishes the existence of rare oxygen-producing organisms by that period.

Oceans of the Archean Eon (4.0 to 2.5 billion years ago) contained much volcanic-derived ferrous iron (Fe^{2+}), which was deposited as hematite (Fe_2O_3) in BIFs. The oxygen that combined the ferrous iron was provided as a waste product of cyanobacterial metabolism. A major burst in the deposition of BIFs from 3.1 billion to 2.5 billion years ago—peaking about 2.7 billion years ago—cleared the oceans of ferrous iron. This enabled the atmospheric oxygen level to increase appreciably. By the time of the widespread appearance of eukaryotes at 1.8 billion years ago, oxygen concentration had risen to 10 percent of present atmospheric level (PAL). These relatively high concentrations were sufficient for oxidative weathering to take place, as evidenced by hematite-rich fossil soils (paleosols) and red beds (sandstones with hematite-coated quartz grains). A second major peak, which raised atmospheric oxygen levels to 50 percent PAL, was reached by 600 million years ago. It was denoted by the first appearance of animal life (metazoans) requiring sufficient oxygen for the production of collagen and the subsequent formation of skeletons. Furthermore, in the stratosphere during the Precambrian, free oxygen began to form a layer of ozone (O_3), which currently acts as a protective shield against the Sun's ultraviolet rays.

The Development of the Ocean

The origin of Earth's oceans occurred earlier than that of the oldest sedimentary rocks. The 3.85-billion-year-old sediments at Isua in western Greenland contain BIFs that were deposited in water. These sediments, which include abraded detrital zircon grains that indicate water transport, are interbedded with basaltic lavas with pillow structures that form when lavas are extruded under water. The stability of liquid water (that is, its continuous presence on Earth) implies that surface seawater temperatures were similar to those of the present.

Differences in the chemical composition of Archean and Proterozoic sedimentary rocks point to two different mechanisms for controlling seawater composition between the two Precambrian eons. During the Archean, seawater composition was primarily influenced by the pumping of water through basaltic oceanic crust, such as occurs today at oceanic spreading centres. In contrast, during the Proterozoic, the controlling factor was river discharge off stable continental margins, which first developed after 2.5 billion years ago. The present-day oceans maintain their salinity levels by a balance between salts delivered by freshwater runoff from the continents and the deposition of minerals from seawater.

Climatic Conditions

A major factor controlling the climate during the Precambrian was the tectonic arrangement of continents. At times of supercontinent formation (at 2.5 billion, 2.1 to 1.8 billion, and 1.0 billion to 900 million years ago), the total number of volcanoes was limited. There were few island arcs (long, curved island chains associated with intense volcanic and seismic activity), and the overall length of oceanic spreading ridges was relatively short. This relative shortage of volcanoes resulted in low emissions of the greenhouse gas carbon dioxide (CO_2). This contributed to low surface temperatures and extensive glaciations. In contrast, at times of continental breakup, which led to maximum rates of seafloor spreading and subduction (at 2.3 to 1.8 billion, 1.7 to 1.2 billion, and 800 to 500 million years ago), there were high emissions of CO_2 from numerous volcanoes in oceanic ridges and island arcs. The atmospheric greenhouse effect was enhanced, warming Earth's surface, and glaciation was absent. These latter conditions also applied to the Archean Eon prior to the formation of continents.

Temperature and Rainfall

The discovery of 3.85-billion-year-old marine sediments and pillow lavas in Greenland indicates the existence of liquid water and implies a surface temperature above 0 °C (32 °F) during the early part of Precambrian time. The presence of 3.5-billion-year-old stromatolites in Australia suggests a surface temperature of about 7 °C (45 °F). Extreme greenhouse conditions in the Archean caused by elevated atmospheric levels of carbon dioxide from intense volcanism (effusion of lava from submarine fissures) kept surface temperatures high enough for the evolution of life. They counteracted the reduced solar luminosity (rate of total energy output from the Sun), which ranged from 70 to 80 percent of the present value. Without these extreme greenhouse conditions, liquid water would not have occurred on the Earth's surface.

In contrast, direct evidence of rainfall in the geological record is very difficult to find. Some limited evidence has been provided by well-preserved rain pits in 1.8-billion-year-old rocks in southwestern Greenland.

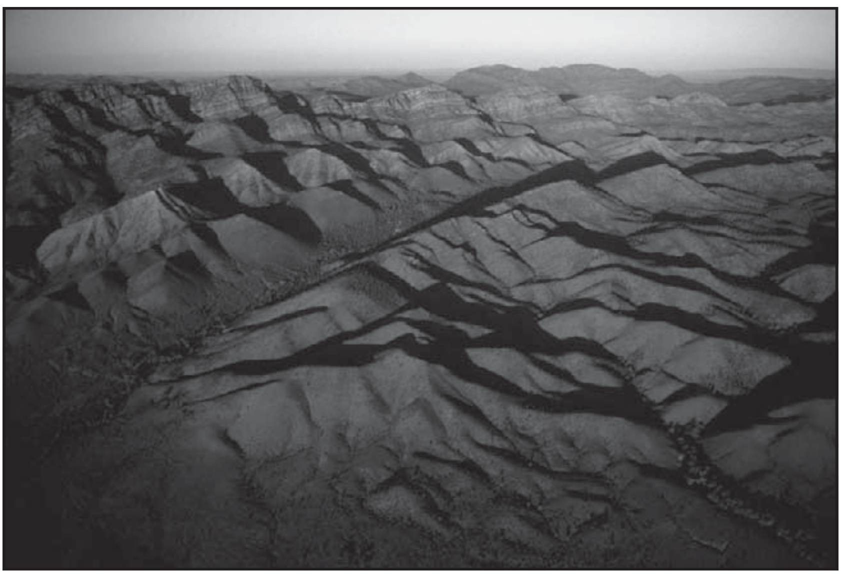
Worldwide Glaciations

The presence of tillites (glacial sediments) indicates that extensive glaciations occurred several times during the Precambrian. Glacial deposits are not necessarily limited to high latitudes. In general, they are complementary to the carbonates, evaporites, and red beds that are climatically sensitive and restricted to low latitudes.

The oldest known glaciation took place 2.9 billion years ago in South Africa during the Late Archean; the evidence is provided by glacial deposits in sediments of the Pongola Rift in southern Africa. The most extensive early Precambrian Huronian glaciation occurred 2.3 billion years ago during the early Proterozoic. It can be recognized from the rocks and structures that the glaciers and ice sheets left behind in parts of Western Australia, Finland, southern Africa, and North America. The most extensive occurrences are found in North America in a belt nearly 3,000 km (1,800 miles) long extending from Chibougamau in Quebec through Ontario to Michigan and southwestward to the Medicine Bow Mountains of Wyoming. This probably represents the area of the original ice sheet. Most details are known from the Gowganda Formation in Ontario, which contains glacial deposits that are up to 3,000 metres (9,850 feet) thick and that occupy an area of about 20,000 square km (7,700 square miles); the entire glacial event may have covered an area of more than 2.5 million square km.

Paleomagnetic studies indicate that the Gowganda glaciation occurred near the paleoequator. Similar, roughly contemporaneous glacial deposits can be found in other parts of the world, suggesting that there was at least one extensive glaciation during the early Proterozoic.

The largest glaciation in the history of the Earth occurred during the late Proterozoic in the period between 1 billion and 600 million years ago. It left its mark almost everywhere. One of the best-described occurrences is in the Flinders Range of South Australia, where there is a sequence 4 km (2.5 miles) thick of tillites and varved sediments occupying an area of 400 by 500 km (250 by 300 miles). Detailed stratigraphy and isotopic dating show that three worldwide glaciations took place: the Sturtian glaciation (750 to 700 million years ago), the Varanger-Marinoan ice ages (625 to 580 million years ago), and the Sinian glaciation (600 to 550 million years ago).



The softly curved mountains of South Australia's Flinders Range have been eroded over the past 500 million years, revealing quartzite at the mountains' highest peaks. Jason Edwards/National Geographic/Getty Images

What is the explanation for all these occurrences of glacial deposits? Some paleomagnetic studies have shown that the tillites in Scotland, Norway, Greenland, central Africa, North America, and South Australia were deposited in low or near-equatorial paleolatitudes. Such conclusions are, however, controversial, because it has also been suggested that the positions of the northern and southern magnetic poles may have migrated across the globe, leaving a record of glaciations in both high and low latitudes. There is the possibility that floating ice sheets could have traveled to low latitudes, depositing glacial sediments and dropstones below them. Whatever the answer, the existence of such vast quantities of tillites and of such extensive glaciations is intriguing. It has been suggested that they record the existence of a frozen "snowball" or "slushball" Earth.

PRECAMBRIAN LIFE

Precambrian rocks were originally defined to predate the Cambrian Period and therefore all life, although the term Proterozoic was later coined from the Greek for "early life." It is now known that Precambrian rocks contain evidence of the very beginnings of life on Earth (and thus the record of its evolution for more than 3.5 billion years), the explosion of life-forms without skeletons before the Cambrian, and even the development of sexual reproduction.

The earliest signs of life on Earth are in western Greenland where apatite (calcium phosphate) grains within a 3.85-billion-year-old meta-sedimentary rock have carbon isotope ratios that indicate an organic origin. The presence of organic hydrocarbon droplets in kerogenous sediments has been found in the 3.46-billion-year-old

Warrawoona Group in the Pilbara craton of Western Australia. These are small amounts of Archean oil.

The first fossil evidence of terrestrial life is found in the early Archean sedimentary rocks of the greenstone-granite belts (metamorphosed oceanic crust and island arc complexes) of the Barberton craton in South Africa and in the Warrawoona Group, which are both about 3.5 billion years old. There are two types of these early, simple, biological structures: microfossils and stromatolites (sheetlike mats precipitated by communities of microorganisms).

Microfossils and Stromatolites

The microfossils occur in cherts and shales and are of two varieties. One type consists of spherical carbonaceous aggregates, or spheroids, which may measure as much as 20 mm (0.8 inch) in diameter. These resemble algae and cysts of flagellates and are widely regarded as biogenic (produced by living organisms). The other variety of microfossils is made up of carbonaceous filamentous threads, which are curving, hollow tubes up to 150 micrometres (0.006 inch) long. Most likely, these tubes are the fossil remains of filamentous organisms. Hundreds of them have been found in some rock layers. The 2.8-billion-year-old gold reefs (conglomerate beds with rich gold deposits) of the Witwatersrand Basin in South Africa contain carbonaceous columnar microfossils up to 7 mm (slightly less than 0.3 inch) long that resemble modern algae, fungi, and lichens. They probably extracted gold from their environment in much the same way that modern fungi and lichens do.

Stromatolites are stratiform, domal, or columnar structures made from sheetlike mats precipitated by communities of microorganisms, particularly filamentous blue-green algae. The early Archean examples form domes as tall as about 10 cm (4 inches). Stromatolites occur in many of the world's greenstone-granite belts. In the 2.7-billion-year-old Steep Rock Lake belt in Ontario, Can., they reach 3 metres (9 feet) in height and diameter. Stromatolites continued to form all the way through the geologic record and today grow in warm intertidal waters, as exemplified by those of Shark Bay in Western Australia. They provide indisputable evidence that life had begun on Earth using algal photosynthesis in complex, integrated biological communities by 3.5 billion years ago.

These Archean organisms were prokaryotes that were incapable of cell division. They were relatively resistant to ultraviolet radiation and thus were able to survive during Earth's early history when the atmosphere lacked an ozone layer. The prokaryotes were predominant until about 1.7 billion to 1.9 billion years ago, when they were overtaken by the eukaryotes (organisms possessing nucleated cells). The latter made use of oxygen in metabolism and for growth and thus developed profusely in the increasingly oxygen-rich atmosphere of the early Proterozoic. The eukaryotes were capable of cell division, which allowed DNA (deoxyribonucleic acid), the genetic coding material, to be passed on to succeeding generations.

By early Proterozoic time, both microfossils and stromatolites had proliferated. The best-known occurrence of microorganisms is in the 2-billion-year-old, stromatolite-bearing Gunflint iron formation in the Huronian Basin of southern Ontario. These microbial fossils include some 30 different types with spheroidal, filamentous, and spore-like forms up to about 20 micrometres (0.0008 inch) across. Sixteen species in 14 genera have been classified so far. Microfossils of this kind are abundant, contain beautifully preserved organic matter, and are extremely similar to such present-day microorganisms as blue-green algae and microbacteria. There are comparable microfossils from the early Proterozoic in Minnesota and Michigan in the United States, the Belcher Islands in Hudson Bay in Canada, southern Greenland, Western Australia, and northern China. These microbiota lived at the time of the transition in the chemical composition of the atmosphere when oxygen began accumulating for the first time.

During the late Proterozoic, stromatolites reached their peak of development, became distributed worldwide, and diversified into complex, branching forms. From about 700 million years ago, however, they began to decline significantly in number. Possibly the newly arrived metazoans (multicelled organisms whose cells are differentiated into tissues and organs) ate the stromatolitic algae, and their profuse growth destroyed the habitats of the latter.

There is the intriguing question as to when sexual division arose in life-forms. In the late 1960s, American paleobiologist J. William Schopf pointed out that the abundant microflora of the 900-million-year-old Bitter Springs Formation of central Australia includes some eukaryotic algae that have cells in various stages of

BITTER SPRINGS MICROFOSSILS

This assemblage of microscopic fossil structures was uncovered in the Bitter Springs Formation, a rock layer about 800 million years old exposed in central Australia. Collections first made in 1965 revealed at least four general groups of organisms that possibly inhabited shallow seas of central Australia in Late Precambrian times (ending about 540 million years ago). These groups resemble bacteria, filamentous blue-green algae, green algae, and fungi. The demonstration of cell division in a fossil green alga named *Glenobotrydion* is evidence that an evolutionary stage that would later lead to sexual reproduction and genetic variation already had been attained.

division arranged into tetrahedral sporelike forms. These resemble the tetrad of spore cells of living plants known to develop by sexual division. In effect, by the end of the Precambrian, the conditions were set for the explosion of life at the start of the Phanerozoic Eon.

GUNFLINT MICROFOSSILS

This assemblage of microscopic fossils was uncovered in the Gunflint Iron Formation, a rock layer about two billion years old exposed in western Ontario, Canada. The fossils include filamentous structures resembling blue-green algae (such as *Gunflintia*, *Entosphaeroides*, and *Animikiea*), tiny spheroids (such as *Eosphaera* and *Huroniospora*), star-shaped forms assigned to the genus *Eoastrion*, and umbrella-shaped forms assigned to the genus *Kakabekia.* These and other fossils were first collected near Thunder Bay, Ont., in the 1950s. Analyses yield strong evidence that some of these fossils are the remains of some of the earliest photosynthetic organisms.

Ediacaran Fossils

Metazoans developed rapidly from the beginning of the Cambrian, when organisms acquired the ability to produce the protein collagen and, thus, skeletons and shells. However, more-primitive metazoans without skeletons—the Ediacara fauna—appeared earlier (more than 600 million years ago), after the end of the Varanger-Marinoan ice age at 580 million years ago and before the onset of the Cambrian Period at 542 million years ago. Ediacaran fossils have been deposited in environments ranging from tidal marine habitats to the deep seafloor. The Ediacaran organisms were probably the ancestors of shelled organisms that mark the beginning of the Phanerozoic.

EDIACARA FAUNA

This unique assemblage of soft-bodied organisms was preserved worldwide as fossil impressions in sandstone from the Proterozoic Eon at the close of Precambrian time. The Ediacara fauna, named for the Ediacara Hills of South Australia in which they were discovered in 1946, were the first metazoans (animals made up of more than one type of cell) that required atmospheric oxygen for their growth. These animals were the precursors of organisms with skeletons, the appearance of which marked the end of Precambrian time and the beginning of the Phanerozoic Eon. The discovery of the Ediacara fauna demonstrated that a far more complex level of evolution had been achieved during Precambrian time than had been previously thought.

The fossil impressions of the Ediacara fauna have a wide variety of shapes, ranging from circular discs (made up of internal radial arrangements, concentric ribbed structures, or combinations) and amorphous masses to plantlike fronds. The disc-shaped impressions are commonly a few centimetres across, though rare specimens reach 20 cm (almost 8 inches) in diameter. The frond-shaped impressions can attain lengths of about 1 metre (3 feet). Accompanying all types of impressions are "trace fossils"—irregularly shaped, winding burrows on and beneath the surface of the sandstone beds.

The Ediacara impressions were derived from soft-bodied organisms similar to modern-day jellyfish, lichen, soft corals, sea anemones, sea pens, annelid worms, and seaweed, as well as some organisms unlike any that are known today. They lived on or near the surface of coarse-grained sediments in the shallow continental shelf or on the deep continental slope of late Precambrian continental margins. Ediacara remains occur in rocks ranging in age from approximately 600 million to 542 million years old; the most-complex forms occur in the last 20 million years of this interval. The oldest radiometrically dated assemblage of Ediacaran organisms in the world, found in the Avalon Zone of Newfoundland, has an age of 565 million years.

It had long been thought that the Ediacara fauna became entirely extinct at the end of the Precambrian, most likely due to heavy grazing by early skeletal animals. However, more recently, it was thought that environmental events such as changes in sea level played a greater role in the extinction of many Ediacaran organisms. Yet, recent discoveries have led to the current view that a few Ediacara-type organisms continued into the Cambrian. Moreover, some calcareous shelly fossils and sponge spicules have been found in Ediacara-age sediments, indicating that there was some degree of diachronous transition between the Precambrian soft-bodied organisms and the organisms with skeletons in the Cambrian.

Most of the Ediacara fauna are found immediately above tillites (glacial beds derived from ice sheets) that were widespread in the late Precambrian. Though it has been suggested that development of the Ediacara organisms was aided by the improvement of climate after the ice ages, a few occurrences of the Ediacara fauna are located between two glacial tillite beds, and some in West Africa and northwestern Canada have been found immediately below a layer of tillites. It is more likely that the origin of the Ediacara fauna was related to a global rise in the atmospheric oxygen level, which triggered a burst of development in these primitive metazoan animals toward the end of Precambrian time.

Fossils of Ediacara organisms have been discovered in some 30 localities over five continents, including seven sites in North America. The principal occurrence is in South Australia's Ediacara Hills, which are part of the Flinders Range and are located 650 km (about 400 miles) north of Adelaide. More than 60 species have been defined from the fossils contained in the Pound Quartzite formation at this site. Other important sites are in central England (Charnwood Forest in Leicestershire), southeastern Newfoundland, northwestern Canada (the Mackenzie and Wernecke mountains), Namibia, Iran, Ukraine, the White Sea, the Urals, northern Siberia, the Yangtze valley in China, and several localities in central Australia.

PRECAMBRIAN GEOLOGY

Rocks laid down during the Precambrian have been dated to one of up to three eons of geologic time, and the oldest rocks and minerals date to over four billion years ago. Mountain-building events during the Archean Eon produced belts of greenstone-granite and granulite-gneiss, which formed the foundations of Earth's continents. As

This folded belt of Archean gneiss can be found in Wyoming. Dr. Marli Miller/Visuals Unlimited/Getty Images

these large regions of rock collided with one another during the Proterozoic, new orogenic belts, sedimentary sequences, and intrusions of igneous rock appeared. Later, as glaciers advanced across Proterozoic continents, they distributed tillites to many parts of the world. Much of the world's stock of precious metal owes its origin to the geologic activities that occurred during Precambrian time.

The Major Subdivisions of the Precambrian System

By international agreement, Precambrian time is divided into the Archean Eon (occurring between roughly 4.0 billion years ago and 2.5 billion years ago) and Proterozoic Eon (occurring between 2.5 billion and 542 million years ago). Many scientists also associate an informal interval, the Hadean Eon (occurring between 4.6 billion and 4.0

billion years ago), with Precambrian time. After the Precambrian, geologic time intervals are commonly subdivided on the basis of the fossil record. The paucity of Precambrian fossils, however, precludes the creation of small-scale subdivisions (epochs and ages) in this time period. Instead, relative chronologies of events have been produced for different regions based on such field relationships as unconformities (interruption in the accumulation of sedimentary rock due to erosion or nondeposition) and crosscutting dikes (intrusions of igneous rock that burrow through cracks in the original structures of surrounding rock). These field relationships, combined with the isotopic age determinations of specific rocks, allow for some correlation between neighbouring regions. The International Commission on Stratigraphy (ISC) and International Union of Geological Sciences (IUGS) divide the Archean Eon into the Eoarchean (approximately 4.0 billion to 3.6 billion years ago), Paleoarchean (3.6 billion to 3.2 billion years ago), Mesoarchean (3.2 billion to 2.8 billion years ago), and Neoarchean (2.8 billion to 2.5 billion years ago) eras. Likewise, they divide the Proterozoic Eon into the Paleoproterozoic (2.5 billion to 1.6 billion years ago), Mesoproterozoic (1.6 billion to 1 billion years ago), and Neoproterozoic (1 billion to 542 million years ago) eras. These definitions are based on isotopic age determinations.

The Oldest Minerals and Rocks

The oldest minerals on Earth, detrital zircons from western Australia, crystallized about 4.4 billion years ago. They occur within sedimentary sandstones and conglomerates dated to about 3.3 billion years ago, but the environment in which they were formed is totally unknown. The rocks from which they came may have been destroyed by some kind of tectonic process or by a meteorite impact that spared individual zircon crystals. On the other hand, rocks containing these minerals may still exist on Earth's surface but simply have not been found. Perhaps their very absence is indicative of something important about early terrestrial processes. Comparisons with the Moon indicate that the Earth must have been subjected to an enormous number of meteorite impacts about 4 billion years ago, but there is no geologic evidence of such events.

The oldest known rocks on Earth are the faux amphibolite volcanic deposits of the Nuvvuagittuq greenstone belt in Quebec, Canada. They are estimated to be 4.28 billion years old. The age of these rocks was estimated using a radiometric dating technique that measures the ratio of the rare-earth elements neodymium and samarium present in a sample.

Precambrian bedrock of the Canadian Shield rising out of Reindeer Lake, on the border between northeastern Saskatchewan and northwestern Manitoba. © Richard Alexander Cooke III

The Acasta gneisses, found near Canada's Great Slave Lake, are also among the world's oldest rocks. Their age has been established radiometrically at 4.0 to 3.9 billion years. The Acasta gneisses are granitic and contain a single relict zircon crystal, which has been dated to 4.2 billion years ago and formed from granitic magma. They are thought to have evolved from older basaltic material in the crust that was melted and remelted by tectonic processes.

Shown here is an Acasta gneiss rock sample from northwest Canada. SSPL via Getty Images

Significant Geologic Events

The Archean and Proterozoic eons within Precambrian time are very different and must be considered separately. The Archean-Proterozoic boundary constitutes a major turning point in Earth history. Before that time the crust of the Earth was in the process of growing, and so there were no large, stable continents. Afterward, when such continents had emerged, orogenic belts were able to form on the margins of and between continental blocks.

There are two types of Archean orogenic belts. The first occurs in upper crustal greenstone-granite belts rich in volcanic rocks that are probably primitive types of oceanic crust and island arcs (long, curved island chains associated with intense volcanic and seismic activity) that formed during the early rapid stage of crustal growth. The second occurs in granulite-gneiss belts that were recrystallized in the Archean mid-lower crust under metamorphic conditions associated with high-temperature granulite and amphibolite facies. Thus, granulites, which typically contain the high-temperature mineral hypersthene (a type of pyroxene), are a characteristic feature of many Precambrian orogenic belts that have been deeply eroded. In Phanerozoic orogenic belts, granulites are rare.

There are several other rock types that developed primarily during the Precambrian but rarely later. This restriction is a result of the unique conditions that prevailed during Precambrian time. For example, banded-iron formations are ferruginous sediments that were deposited on the margins of early, iron-rich oceans. Anorthosite, which consists largely of plagioclase, forms large bodies in several Proterozoic belts. Komatiite, a magnesium-rich, high-temperature volcanic rock derived from very hot mantle (part of the Earth between the crust and the core), was extruded in abundance during the early Precambrian when the heat flow of the Earth was higher than it is today. Blueschist, which contains the blue mineral glaucophane, forms in subduction zones under high pressures and low temperatures, and its rare occurrence in Precambrian rocks may indicate that temperatures in early subduction zones were too high for its formation.

The bulk of many of the world's valuable mineral deposits (for example, those of gold, nickel, chromite, copper, and iron) also formed during the Precambrian. These concentrations are a reflection of distinctive Precambrian sedimentary and magmatic rocks and their environments of formation.

Archean Crustal Growth

During the first third of geologic history (that is, until about 2.5 billion years ago), the Earth developed in a broadly similar manner. Greenstone-granite belts (metamorphosed oceanic crust and island arc complexes) formed in the upper Archean crust, and granulite-gneiss belts formed in the mid-lower crust. This was a time when the overall rate of heat production by the breakdown of radioactive isotopes was several times greater than it is today. This condition was manifested by very rapid tectonic processes, probably by some sort of primitive plate tectonics (more-modern plate-tectonic processes could not occur until the crust became cooler and more rigid). Most of the heat that escapes from Earth's interior today does so at oceanic ridges. This manner of heat loss probably occurred during the Archean in much larger amounts. The oceanic ridges of the Archean were more abundant, longer, and opened faster than those in the modern oceans, and oceanic plateaus derived from hot mantle plumes (slowly rising currents of highly viscous mantle material) were more common. Although the amount of newly generated crust was probably enormous, a large part of this material was inevitably destroyed by equally rapid plate subduction processes. The main results of this early growth that still remain today are the many island arcs and oceanic plateaus in greenstone-granite belts and the voluminous Andean-type tonalites (a granitic-type rock rich in plagioclase feldspar) that were deformed to orthogneiss (gneiss derived from igneous rocks) in granulite-gneiss belts. Although most of the Archean oceanic crust was subducted, a few ophiolitic-type complexes have been preserved in greenstone-granite belts.

The late Archean (Neoarchean Era) was an important interval of time because it marks the beginning of the major changeover from Archean to Proterozoic types of crustal growth. The formation of the first major rifts characterized the significant events of this time. The first major rift valley known in the world, the Pongola Rift, emerged along the border of present-day Swaziland and South Africa. The intrusion of the first major basic dikes (such as the Great Dyke, which transects the entire Zimbabwe craton) and the first large stratiform layered igneous complexes (such as the Stillwater in Montana) formed; and the formation of the first large sedimentary basins (for example, the Witwatersrand in South Africa) also occurred. All of these structures indicate that the continental crust had reached a mature stage with considerable stability and rigidity for the first time during the late Archean. The Neoarchean represents the culmination that followed the rapid tectonic processes of the early Archean (Eoarchean and Paleoarchean) and middle Archean (Mesoarchean) eras. Because crustal growth took place at different times throughout the world, similar structures can be found in the early Proterozoic (Paleoproterozoic) Era.

Proterozoic Plate Movements

During the early Proterozoic, large amounts of quartzite, carbonate, and shale were deposited on the shelves and margins of many continental blocks. This would be consistent with the breakup of a supercontinent into several smaller continents with long continental margins (combined areas of continental shelf and continental slope). Examples of shelf sequences of this kind are found along the margins of orogenic (mountain) belts, such as the Wopmay, bordering Canada's Slave province, and also the Labrador Trough, bordering the Superior province.

The existence of stable continental blocks by the early Proterozoic allowed orogenic belts to develop at their margins by some form of collision tectonics. This was the first time that long, linear orogenic belts could form by "modern" tectonic processes that involved seafloor spreading, ophiolite obduction, subduction, and landmass collisions. Subduction lead to the creation of island arcs and Andean-type (formed by subduction at the continental margin) granitic batholiths. In addition, the collision of arcs and continents could now give rise to both sutures with ophiolites and to Himalayan-type (formed by continent-to-continent collision) thrust belts with abundant crustal-melt granites. These were key events in the evolution of the continents, and such processes have continued throughout Earth history.

During the late Proterozoic (Neoproterozoic Era), some orogenic belts, like the Pan-African belts of Saudi Arabia and East Africa, continued to develop. The intense crustal growth and the many orogenic belts that formed throughout the Proterozoic began to create large continental blocks, which amalgamated to produce a new supercontinent by the end of the Precambrian. Therefore, in the late Proterozoic many sedimentary basins were infilled with conglomerates and sandstones due to the deposition of material eroded from higher elevations. For example, the Riphean sequence in Russia and also the Sinian sequence in China were able to form on extensive cratons of continental crust.

The Occurrence and Distribution of Precambrian Rocks

Precambrian rocks, as a whole, occur in a wide variety of shapes and sizes. There are extensive Archean regions, up to a few thousands of kilometres across, that may contain either greenstone-granite belts or granulite-gneiss belts or both. These regions are variously designated in different parts of the world as cratons, shields, provinces, or blocks. Some examples include: the North Atlantic craton that incorporates northwestern Scotland, central Greenland, and Labrador; the Kaapvaal and Zimbabwean cratons in southern Africa; the Dharwar craton in India; the Aldan and Anabar shields in Siberia in Russia; the Baltic Shield that includes much of Sweden, Finland, and the Kola Peninsula of far northern Russia; the Superior and Slave provinces in Canada; and the Yilgarn and Pilbara blocks in Western Australia. Linear belts, up to several thousand kilometres long, that are frequently though not exclusively of Proterozoic age include the Limpopo, Mozambique, and Damaran belts in Africa, the Labrador Trough in Canada, and the Eastern Ghats belt in India. Several small relict areas, spanning a few hundred kilometres across, exist within or against Phanerozoic orogenic belts. These include the Lofoten islands of Norway, the Lewisian Complex in northwestern Scotland, and the Adirondack Mountains in the northeastern United States. Nevertheless, some extensive areas of Precambrian rocks—such as under the European and Russian platforms and under the central United States—remain overlain by a blanket of Phanerozoic sediments.

Archean Rock Types

Archean rocks occur in greenstone-granite belts that represent the upper crust, in granulite-gneiss belts that formed in the mid-lower crust, and in sedimentary basins, basic dikes, and layered complexes that were either deposited on or intruded into the first two types of belts.

Greenstone-Granite Belts

These belts occur on most continents. The largest extend several hundred kilometres in length and measure several hundred metres in width. Today, many greenstone-granite belts are regarded as tectonic "slices" of oceanic and island arc crust that have been thrust together to form tectonic collages similar to those in belts found in the present-day Pacific Ocean.

The greenstone sequence in many belts is divisible into a lower volcanic group and an upper sedimentary group. The volcanics are made up of lavas that are ultramafic (silica content less than 45 percent) and basaltic (silica content of 45 to 52 percent). The uppermost sediments are typically terrigenous (land-derived) shales, sandstones, quartzites, wackes, and conglomerates. All the greenstone sequences have undergone recrystallization during the metamorphism of greenschist facies at relatively low temperatures and pressures. In fact, the presence of the three green metamorphic minerals chlorite, hornblende, and epidote has given rise to the term *greenstone* for the recrystallized basaltic volcanics. Granitic rocks and gneisses occur within, adjacent to, and between many greenstone sequences.

The Economic Significance of Archean Greenstone-Granite Deposits

Abundant mineralization has occurred in greenstone-granite belts. These belts constitute one of the world's principal depositories of gold, silver, chromium, nickel, copper, and zinc. In the past they were termed gold belts because of the gold rushes of the 19th century that took place in areas such as Kalgoorlie in the Yilgarn belt of Western Australia, the Barberton belt of South Africa, and Val d'Or in the Abitibi belt of southern Canada. The mineral deposits occur in all the major rock groups: chromite, nickel, asbestos, magnesite, and talc in ultramafic lavas; gold, silver, copper, and zinc in basaltic to rhyolitic volcanics; iron ore, manganese, and barite in sediments; and lithium, tantalum, beryllium, tin, molybdenum, and bismuth in granites and associated pegmatites. Important occurrences are chromite at Selukwe in Zimbabwe, nickel at Kambalda in southwestern Australia, tantalum in Manitoba in Canada, and copper-zinc at Timmins and Noranda in the Canadian Abitibi belt.

Greenstone-Granite Rock Types

The volcanics that comprise the lower portion of a greenstone sequence are made up of lavas noted for magnesian komatiites (ultramafic extrusive igneous rocks) that probably formed in the oceanic crust that are overlain by basalts, andesites, and rhyolites whose chemical composition is much like that of modern island arcs. Especially important is the presence in the Isua, Barberton, and Yellowknife belts of sheeted basic dike complexes cutting across gabbros and overlain by pillow-bearing basalts (basalts extruded underwater that form characteristic pillow-shaped hummocks). Volcanic sequences are capped by oceanic cherts and terrigenous sedimentary groups. The overall stratigraphy suggests an evolution from extensive submarine eruptions of komatiite and basalt (ocean floor) to more-localized stratovolcanoes (volcanoes constructed from alternating layers of ash and lava), which become increasingly emergent with intervening and overlying clastic sediments (clay-, silt-, and sand-sized sediments) that were deposited in trenches at the mouths of subduction zones. There are, however, regional differences in the volcanic and sedimentary makeup of some belts. The older belts in southern Africa and Australia have more komatiites, basalts, shallow-water banded-iron formations, cherts, and evaporites and fewer terrigenous (land-derived) sediments. On the other hand, the younger belts in North America have a higher proportion of andesites, rhyolites, and terrigenous and turbidite debris (sediments delivered to the deep ocean by density currents) but fewer shallow-water sediments. These differences reflect a change from the older oceanic-type volcanism (effusion of lava from submarine fissures) to the younger, more arc-type phenomena such as explosive eruption of pyroclastic materials (incandescent material ejected during violent eruptions) and lava flows from steep volcanic cones. Additional changes include an increase in the amount of trench (subduction zone) turbidites and graywackes and an increase in the availability of continental crust as a source for terrigenous debris.

Ultramafic rocks (rocks with a very low silica content—less than 45 percent) are commonly altered to talc schists and tremolite-actinolite schists. There are some indications that several phases of metamorphism exist—namely, seafloor metamorphism associated with the action of hydrothermal brines that could occur at oceanic ridges, syntectonic metamorphism related to thrust-nappe tectonics, and local thermal contact metamorphism caused by intrusive granitic plutons pushing into cooler surrounding rock.

Granitic rocks and gneisses are associated with many greenstone sequences. Some paragneisses (gneisses metamorphosed from sedimentary rocks), as in the Quetico belt in Canada, are derived from wackes. They were probably deposited in an ocean trench or accretionary prism (a mass of accumulating sediments on the inner trench wall in a subduction zone) at the mouth of a subduction zone between the island arcs of the adjacent greenstone sequences. Many early granitic plutons were deformed and converted into orthogneiss (gneisses metamorphosed from igneous rocks). Late plutons commonly intruded the greenstones that were downfolded in synclines (an upward concave fold of rock) between them, or they intruded along the borders of the belts, deflecting them into irregular shapes.

The Structure and Formation of Greenstone-Granite Belts

The structure of many belts is complex. Their stratigraphic successions are upside-down and deformed by thrusts and major horizontal folds (nappes). They have been subsequently refolded by upright anticlines (convex folds of rock) and synclines. The result of this thrusting is the repetition of the same stratigraphic successions on top of one another, creating a massive deposit of material up to 10 to 20 km (6 to 12 miles) thick. Also, there may be thrusts along the base of the belts, as in the case of Barberton, showing that they have been transported from elsewhere. In other instances, the thrusts may occur along the borders of the belts, indicating that they have been forced against and over adjacent gneissic belts. The conclusion from structural studies is that many belts have undergone intense subhorizontal deformation during thrust transport.

Clearly, there are different types of greenstone-granite belts. To understand their origin and mode of evolution, it is necessary to correlate them with comparable modern analogues. Some, like the Barberton and Yellowknife belts, consist of oceanic-type crust and have sheeted dike swarms that occur in many ophiolites of Mesozoic-Cenozoic origin, such as in the Troodos Mountains in Cyprus. They are the hallmark of a modern oceanic crust that formed at an oceanic ridge. Also, like modern ophiolites, a few seem to have been covered by thrusting onto continental crust. Many belts, such as the Isua belt of Greenland and those in the Superior province of Canada, are very similar to modern island arcs. Geochemical data are revealing that some lavas were derived from depths of 1,000 to 2,700 km (620 to 1,680 miles) in the Earth's mantle and not from shallower subduction zones, which are commonly 600 km (about 373 miles) deep. These rocks are comparable to oceanic plateaus in modern oceanic crust that were formed from plumes of hot magma from the very deep mantle. The Wawa belt, for example, has been shown to consist of an immature island arc built on oceanic plateau crust and overlain by a more mature arc. The Abitibi belt began as oceanic crust with island arcs and oceanic plateaus. Between the Wawa and Wabigoon island arcs lies the Quetico belt, consisting of metamorphosed turbidites and slices of volcanics that probably developed in a regularly overlapping accretionary prism in an arc-trench system, as seen today in the Japanese arcs. The Pilbara belts are similar to modern active continental margins, and they have been interthrust with older continental orthogneisses to form very thick crustal piles intruded by diapiric crustal-melt granites. This scenario is quite comparable to that of a Himalayan type of orogenic belt formed by collisional tectonics. In conclusion, most greenstone-granite belts are today regarded by geologists as different parts of interthrust oceanic crust-accretionary prism structures within island arcs of oceanic plateau systems that collided with continental gneissic blocks.

The Age and Occurrence of Greenstone-Granite Belts

Greenstone-granite belts developed at many different times throughout the long Archean Eon. The Isua greenstone belt in West Greenland is about 3.85 billion years old. In the Zimbabwean craton, they formed over three successive periods: the Selukwe belt about 3.8 to 3.75 billion years ago, the Belingwean belts about 2.9 billion years ago, and the Bulawayan-Shamvaian belts about 2.7 to 2.6 billion years ago. The Barberton belt in the Kaapvaal craton and the Warrawoona belt in the Pilbara block are 3.5 billion years old. Globally, the most important period of formation was from 2.7 to 2.6 billion years ago, especially in the Slave and Superior provinces of North America, the Yilgarn block in Australia, and the Dharwar craton in India. Some of the better-documented belts seem to have formed within about 50 million years. It is important to note that while the Bulawayan-Shamvaian belts were forming in the Zimbabwean craton, flat-lying sediments and volcanics were laid down in the Pongola Rift and the Witwatersrand Basin not far to the north.

Greenstone-granite belts range from aggregates of several belts (as in the southern Superior province of Canada) to irregular, even triangular-shaped belts (as in the Barberton in South Africa) to synclinal basins (as in the Indian Dharwar craton). The irregular and synclinal shapes are commonly caused by the diapiric intrusion of younger granites.

Important occurrences are the Barberton belt in South Africa; the Sebakwian, Belingwean, and Bulawayan-Shamvaian belts of Zimbabwe; the Yellowknife belts in the Slave province of Canada; the Abitibi, Wawa, Wabigoon, and Quetico belts of the Superior province of Canada; the Dharwar belts in India; and the Warrawoona and Yilgarn belts in Australia.

Granulite-Gneiss Belts

The granulites, gneisses, and associated rocks in these belts were metamorphosed to a high grade in deep levels of the Archean crust. Metamorphism occurred at a temperature of 750 to 980 °C (1,380 to 1,800 °F) and at a depth of about 15 to 30 km (9 to 19 miles). These belts, therefore, represent sections of the continents that have been highly uplifted, with the result that the upper crust made up of volcanics, sediments, and granites has been eroded. Accordingly, the granulite-gneiss belts are very different from the greenstone-granite belts. Granulite-gneiss belts may be regarded as variably preserved sections of continental cratons.

The Economic Significance of Archean Granulite-Gneiss Deposits

The mid-lower crust is relatively barren of ore deposits as compared to the upper crust with its sizable concentrations of greenstones and granites, and therefore little mineralization is found in the granulite-gneiss belts. The few exceptions include a nickel-copper sulfide deposit at Selebi-Pikwe in the Limpopo belt in Botswana that is economic to mine, and banded-iron formations in gneisses in the eastern Hubei and Liaoning provinces of northwestern China that form the foundation of a major steel industry. There are subeconomic quantities of chromitite in the anorthosites of western Greenland, southern India, and the Limpopo belt; iron from a banded-iron formation at Isua in western Greenland; and tungsten in amphibolites of western Greenland.

Granulite-Gneiss Rock Types

Orthogneisses of deformed and recrystallized tonalite (a granitic-type rock rich in plagioclase feldspar) and granite constitute the most common rock type. The geochemical signature of these rocks closely resembles that of modern equivalents that occur in granitic batholiths in the Andes. Where such rocks have been metamorphosed under conditions associated with amphibolite facies, they contain hornblende, biotite, or a combination of the two. However, where they have been subjected to conditions of higher temperature associated with the granulite facies, the rocks contain pyroxene and hypersthene and so can be called granulites.

The granulites and gneisses enclose a wide variety of other minor rock types in layers and lenses. These types include schists and paragneisses that were originally deposited on the Earth's surface as shales and which now contain high-temperature metamorphic minerals such as biotite, garnet, cordierite, staurolite, sillimanite, or kyanite. There also are quartzites, which were once sandstones or cherts; marbles (either limestones or dolomites); and banded-iron formations. Commonly intercalated with these metasediments are amphibolites, which locally contain relict pillow structures—demonstrating that they are derived from basaltic lavas extruded underwater. These amphibolites have a trace element chemistry quite similar to that of modern seafloor basalts. The amphibolites are often accompanied by chromite-layered anorthosite, gabbro, and ultramafic rocks such as peridotite and dunite. All these rocks occur in layered igneous complexes, which in their well-preserved state may be up to 2 km (1.2 miles) thick and 100 km (60 miles) long. Such complexes occur at Fiskenaesset in western Greenland, in the Limpopo belt of southern Africa, and in southern India. These complexes may have formed at an oceanic ridge in a magma chamber that also fed the basaltic lavas, or they may be parts of oceanic plateaus. In many cases, the complexes, basaltic amphibolites, and sediments were extensively intruded by the tonalites and granites that were later deformed and recrystallized. The result of this is that all of these rocks may now occur as metre-sized lenses in the orthogneisses and granulites.

The Structure and Occurrence of Granulite-Gneiss Belts

The structure of the granulite-gneiss belts is extremely complex because the constituent rocks have been highly deformed several times. In all likelihood the basalts and layered complexes from the oceanic crust were interthrust with shallow-water limestones, sandstones, and shales; with tonalites and granites from Andean-type batholiths; and with older basement rocks from a continental margin. All these rocks, which are now mutually conformable (parallel to one another with uninterrupted deposition), were folded in horizontal nappes and then refolded. The picture that emerges is one of a very mobile Earth, where newly formed rocks were routinely compressed and thrust against other rocks.

Granulite-gneiss belts occur in a variety of environments. These may be extensive regions, such as the North Atlantic craton—which measures 1,000 by 2,000 km (about 620 by 1,240 miles) across and, before the opening of the Atlantic Ocean, was contiguous with the Scourian Complex of northwestern Scotland—the central part of Greenland, and the coast of Labrador; the Aldan and

Ukrainian shields of continental Europe; the North China craton; large parts of the Superior province of Canada; the Yilgarn block in Australia; and the Limpopo belt in southern Africa. They may be confined to small areas such as the Ancient Gneiss Complex of Swaziland, the Minnesota River valley and the Beartooth Mountains of the United States, the Peninsular gneisses and Sargur supracrustals of southern India, the English River gneisses of Ontario in Canada that form a narrow strip between greenstone-granite belts, the Sand River gneisses that occupy a small area between greenstone-granite belts in Zimbabwe, and the Napier Complex in Enderby Land in Antarctica. Granulite-gneiss belts are commonly surrounded by younger, mostly Proterozoic belts that contain remobilized relicts of the Archean rocks, and the granulites and gneisses must underlie many Archean greenstone-granite belts and blankets of Phanerozoic sediment.

The Age and Correlation of Granulite-Gneiss Belts

Isotopic age determinations from the granulite-gneiss belts record an evolution from about 4.0 to 2.5 billion years ago—more than a third of geologic time. Most important are the few but well-constrained age determinations of detrital zircons at Mount Narryer and Jack Hills in Western Australia that are more than 4 billion years old. Several regions have a history that began in the period dating from 3.9 to 3.6 billion years ago—western Greenland, Labrador, the Limpopo belt, Enderby Land, the North China craton, and the Aldan Shield. Most regions of the world experienced a major tectonic event that may have involved intrusion, metamorphism, and deformation during the period between 3.1 and 2.8 billion years ago. Some of these regions, like the Scourian in northwestern Scotland, show no evidence of any older crustal growth.

The best-documented region is in western Greenland, which has a long and complicated history from 3.85 to 2.5 billion years ago.

It is impossible to correlate the rocks in different granulite-gneiss belts. One granitic gneiss is essentially the same as another but may be of vastly different age. There is a marked similarity in the anorthosites in various belts throughout the world, and their similar relationship with the gneisses suggests that the belts have undergone comparable stages of evolution, although each has its own distinctive features. Little correlation can be made with rocks of Mesozoic-Cenozoic age because few modern orogenic belts have been eroded sufficiently to expose their mid-lower crust. The lack of modern analogues for comparison makes it particularly difficult to interpret the mode of origin and evolution of the Archean granulite-gneiss belts.

Sedimentary Basins, Basic Dikes, and Layered Complexes

During middle and late Archean time (3 to 2.5 billion years ago), relatively stable, post-orogenic conditions developed locally in the upper crust—especially in southern Africa, where the development of greenstone-granite and granulite-gneiss belts was completed much earlier than in other parts of the world. The final chapters of Archean crustal evolution can be followed by considering specific key sedimentary basins, basic (basaltic) dikes, and layered complexes.

Along the border of Swaziland and South Africa is the Pongola Rift, which is the oldest such continental trough in the world. It is 2.95 billion years old, having formed only 50 million years after the thrusting of adjacent greenstone-granite belts. If there were earlier rifts, they have not survived, or, more likely, this was the first time in Earth history that the upper crust was sufficiently stable and rigid for a rift to form. It is 30 km (19 miles) wide, 130 km (81 miles) long, and within it is a sequence of lavas and sediments that is 11 km (7 miles) thick. It seems most likely that the rift developed as the result of the collapse of an overthickened crust following the long period of Archean crustal growth and thrusting in the Kaapvaal craton.

The 200-by-350-km (124-by-217-mile) Witwatersrand Basin contains an 11-km- (7-mile-) thick sequence of lavas and sediments that are 3 billion years old. The basin is

South Africa contains one of the most productive gold mining districts in the world. This gold mine at Wiitwatersrand, near Johannesburg, was photographed in 1955. Evans/Hulton Archive/Getty Images famous for its very large deposits of gold and uranium that occur as detrital minerals in conglomerates. These minerals were derived by erosion of the surrounding greenstone-granite belts and transported by rivers into the shoreline of the basin. In all probability, the gold originally came from the komatiitic and basaltic lavas in the early Archean oceanic crust.

The Great Dyke, thought to be about 2.5 billion years old, transects the entire Zimbabwe craton. It is 480 km (about 300 miles) long, 8 km (5 miles) wide, and made up of layered ultrabasic rocks—gabbros and norites. The ultrabasic rocks have several layers of chromite and an extensive platinum-bearing layer that form economic deposits. The Great Dyke represents a rift that has been filled in with magma that was probably derived from a deep mantle plume.

The Stillwater Complex is a famous, 2.7-billion-year-old, layered ultrabasic-basic intrusion in the Beartooth Mountains of Montana in the United States. It is 48 km (30 miles) long and has a stratigraphic thickness of 6 km (3.7 miles). It was intruded as a subhorizontal body of magma that underwent crystal settling to form the layered structure. It is notable for a 3-metre- (9-foot-) thick layer enriched in platinum minerals that forms a major economic deposit.

The basins, dikes, and complexes described above cannot be mutually correlated. They most resemble equivalent structures that formed at the end of plate-tectonic cycles in the Phanerozoic. They represent the culmination of Archean crustal growth.

Proterozoic Rock Types

What happened geologically at the time of the Archean-Proterozoic boundary 2.5 billion years ago is uncertain. It seems to have been a period of little tectonic activity, and so it is possible that the earlier intensive Archean crustal growth had caused the amalgamation of continental fragments into a supercontinent, perhaps similar to Pangea of Permian-Triassic times. The fragmentation of this supercontinent and the formation of new oceans gave rise to many continental margins upon which a variety of distinctive sediments were deposited. Much evidence suggests that in the period from 2.5 billion to 570 million years ago Proterozoic oceans were formed and destroyed by plate-tectonic processes and that most Proterozoic orogenic belts arose by collisional tectonics. Sedimentary, igneous, and metamorphic rocks that formed during this period are widespread throughout the world. There are many swarms of basic dikes, important sedimentary rifts, basins, and layered igneous complexes, as well as many orogenic belts. The rocks commonly occur in orogenic belts that wrap around the borders of Archean cratons. The characteristic types of Proterozoic rocks are considered below, as are classic examples of their occurrence in orogenic belts. The following types of rocks were formed during the early, middle, and late Proterozoic, indicating that similar conditions and environments existed throughout this long period of time.

Basic Dikes

The continents were sufficiently stable and rigid during the Proterozoic Eon for an extremely large number of basic dikes to be intruded into parallel, extensional fractures in major swarms. Individual dikes measure up to several hundred metres in width and length, and there may be hundreds or even thousands of dikes in a swarm, some having transcontinental dimensions. For example, the 1.2-billion-year-old Mackenzie swarm is more than 500 km (311 miles) wide and 3,000 km (1,864 miles) long and extends in a northwesterly direction across the whole of Canada from the Arctic to the Great Lakes. The 1.95-billion-year-old Kangamiut swarm in western Greenland is only about 250 km (155 miles) long but is one of the world's densest continental dike swarms. Many of the major dike swarms were intruded on the continental margins of Proterozoic oceans in a manner similar to the dikes that border the present-day Atlantic Ocean and were similarly the result of the rise of mantle plumes into the crust.

Layered Igneous Intrusions

There are several very important layered, mafic to ultramafic intrusions of Proterozoic age that were formed by the accumulation of crystals in large magma chambers. The well-known ones are several tens or even hundreds of kilometres across, have a dikelike or sheetlike (stratiform) shape, and contain major economic mineral deposits. The largest and most famous is the Bushveld Complex in South Africa, which is 9 km (5.6 miles) thick and covers an area of 66,000 square km (about 25,500 square miles). It was intruded nearly 2.1 billion years ago and is the largest repository of magmatic ore deposits in the world. The Bushveld Complex consists of stratiform layers of dunite, norite (a type of gabbro rich in orthopyroxene), anorthosite, and ferrodiorite (an iron-rich intrusive igneous rock that is basic to intermediate in composition) and contains deposits of chromite, iron, titanium, vanadium, nickel, and—most important of all—platinum. The Sudbury Complex in southern Canada, which is about 1.9 billion years old, is a basin-shaped body that extends up to 60 km (37 miles) across. It consists mostly of layered norite and has deposits of copper, nickel, cobalt, gold, and platinum. It is noted for its high-pressure structures and other manifestations of shock metamorphism, which suggest that the intrusion was produced by an enormous meteorite impact.

Shelf-Type Sediments

Quartzites, dolomites, shales, and banded-iron formations make up sequences that reach up to 10 km (6.2 miles) in thickness and that amount to more than 60 percent of Proterozoic sediments. Minor sediments include sandstones, conglomerates, red beds, evaporites, and cherts. The quartzites typically have cross-bedding and ripple marks, which are indicative of tidal action, and the dolomites often contain stromatolites similar to those that grow today in intertidal waters. Also present in the dolomites are phosphorites that are similar to those deposited on shallow continental margins against areas of oceanic upwelling during the Phanerozoic. Several early-middle Proterozoic examples of such dolomites have been found in Finland and northern Australia, as well as in the Marquette Range of Michigan in the United States, in the Aravalli Range of Rajasthan in northwestern India, and at Hamersley and Broken Hill in Australia. Other constituents of these dolomites include evaporites that contain casts and relicts of halite, gypsum, and anhydrite. Examples occur at Mount Isa in Australia (1.6 billion years old) and in the Belcher Group in Canada (1.8 billion years old). These evaporites were deposited by brines in very shallow pools such as those encountered today in the Persian Gulf.

Ophiolites

Phanerozoic ophiolites are considered to be fragments of ocean floor that have been trapped between island arcs and continental plates that collided or that have been thrust onto the shelf sediments of continental margins. They consist of a downward sequence of oceanic sediments such as cherts, pillow-bearing basalts, sheeted basic dikes, gabbros, and certain ultramafic rocks (such as serpentinized harzburgite, which is primarily made of olivine and orthopyroxene; and lherzolite, which is mainly composed of olivine, clinopyroxene, and orthopyroxene). Comparable ophiolites occur in several Proterozoic orogenic belts and provide strong evidence of the existence of oceanic plates similar to those of today. The oldest is an ophiolite in the Cape Smith belt on the south side of Hudson Bay in Canada whose age has been firmly established at 1.999 billion years. There is a 1.96-billion-year-old ophiolite in the Svecofennian belt of southern Finland, but most Proterozoic ophiolites are 1 billion to 570 million years old and occur in the Pan-African belts of Saudi Arabia, Egypt, Yemen, and The Sudan, where they occur in sutures between a variety of island arcs.

Greenstones and Granites

Greenstone-granite belts such as those of the Archean continued to form in the Proterozoic, albeit in greatly reduced amounts. They are characterized by abundant volcanic rocks that include pillowed subaqueous basalt flows and subaerial and subaqueous volcaniclastic rocks. Magnesian komatiites are for the most part absent, however. Intrusive plutons are typically made of granodiorite. Examples occur at Flin Flon in central Canada, in the Birrimian Group in West Africa, and in the Pan-African belts of the Arabian-Nubian Shield. Generally, such rocks resemble those in modern island arcs and back-arc basins, and the presence of remnants of oceanic plateau is suspected.

Granulites and Gneisses

These highly deformed and metamorphosed rocks are similar to those of the Archean Eon and occur in many Proterozoic orogenic belts such as the Grenville in

Canada, the Pan-African Mozambique belt in eastern Africa and Madagascar, the Musgrave and Arunta ranges in Australia, and in Lapland in the northern Baltic Shield. They were brought up from the mid-lower crust on major thrusts as a result of continental collisions.

Orogenic Belts

One of the world's classic Proterozoic orogenic belts is the Wopmay Orogen, which is situated in the Arctic in the northwestern part of the Canadian Shield. This beautifully exposed belt formed within a relatively short time (between 1.97 and 1.84 billion years ago) and provides convincing evidence of tectonic activity of a modern form in the early Proterozoic. On the eastern continental margin here are red beds (sandstones) that pass oceanward and westward into stromatolite-rich dolomites deposited on the continental shelf to a thickness of 4 km (2.5 miles). These dolomites pass into submarine turbidite fans that were deposited on the continental rise. An island arc and a continental margin are located to the west. The history of the Wopmay Orogen can be best interpreted in terms of subduction of oceanic crust and collision tectonics.

The Svecofennian Orogen of the Baltic Shield extends in a southeasterly direction from northern Sweden through southern Finland to the adjoining part of western Russia. It formed in the period from 1.9 to 1.7 billion years ago. A major lineament across southern Finland consists of the suture zone on which occur ophiolite complexes representing the remains of oceanic crust. At Outokumpu there is copper mineralization in these oceanic crust rocks similar to that in the Cretaceous ophiolite at Troodos in Cyprus. On the northern side of the suture is a shelf-type sequence of sediments; on the southern side is a volcanic-plutonic arc. To the south of this arc lies a broad zone with thrusted gneisses intruded by tin-bearing crustal-melt granites, called rapakivi granites after their coarse, zoned feldspar megacrysts (that is, crystals that are significantly larger than the surrounding fine-grained matrix). The rocks in this zone probably formed as a result of mantle plume activity.

The Grenville Orogen is a deeply eroded and highly uplifted orogenic belt that extends from Labrador in northeastern Canada to the Adirondack Mountains and southwestward under the coastal plain of the eastern United States. It developed from about 1.5 to 1 billion years ago. Apart from an island arc situated today in Ontario, most of the Grenville Orogen consists of highly metamorphosed and deformed gneisses and granulites that have been brought to the present surface on major thrusts from the mid-lower crust. A result of the terminal continental collision that occurred at about 1.1 billion years ago was the formation of the Midcontinent (or Keweenawan) rift system that extends southward for more than 2,000 km (about 1,240 miles) from Lake Superior.

A type of crustal growth—one very different from that described above—took place in what are now Saudi Arabia, Egypt, Yemen, and The Sudan in the period from 1.1 billion to 500 million years ago. This entire shield, called the Arabian-Nubian Shield, is dominated by volcanic lavas, tuffs (consolidated rocks consisting of pyroclastic fragments and ash), and granitic plutons that formed in a variety of island arcs separated by several sutures along which many ophiolite complexes occur. Some of the ophiolites contain a complete stratigraphy that is widely accepted as a section through the oceanic upper mantle and crust. The final collision of the arcs was associated with widespread thrusting and followed by the intrusion of granitic plutons containing tungsten, tin, uranium, and niobium ore deposits. The island arcs grew from the subduction of oceanic crust in a manner quite comparable to that taking place today throughout Indonesia.

The Mozambique belt is one of the many Pan-African orogenic belts that formed in the period between 1 billion and 500 million years ago. It extends along the eastern border of Africa from Ethiopia to Kenya and Tanzania. It consists largely of highly metamorphosed, mid-crustal gneisses deformed by eastward-dipping thrusts very similar to the thrusts on the southern side of the Himalayas (formed as a result of the collision of India with Tibet during the early Cenozoic Era). To the east on the island of Madagascar, mid-crustal gneisses of similar age were brought to the surface by major late extensional collapse of the orogenic belt.

During the middle and late Proterozoic, thick sequences of sediment were deposited in many basins throughout Asia. The Riphean sequence spans the period from 1.6 billion to 800 million years ago and occurs primarily in Russia. The Sinian sequence in China extends from 800 to 570 million years ago, toward the end of the Precambrian time. The sediments are terrigenous debris characterized by conglomerates, sandstone, siltstone, and shale, some of which are oxidized red beds, along with stromatolite-rich dolomite. Total thicknesses reach over 10 km (6.2 miles). The terrigenous sediments were derived from the erosion of Proterozoic orogenic belts.

Glacial Sediments

Evidence of the oldest known glaciation, which occurred 2.9 billion years ago, is preserved in the Pongola Rift in South Africa, though most Precambrian glaciations occurred during the Proterozoic. Evidence that ancient deposits are of glacial origin is obtained by comparing them with those left behind by the Quaternary ice sheets and with deposits associated with modern glaciers. The main sediments left behind by early Proterozoic glaciers are tillites containing rock fragments ranging in size from pebbles to boulders and distributed randomly in a fine-grained silty matrix. The surfaces of some pebbles have parallel scratches caused by having been rubbed against harder pebbles during ice transport. Locally, the basement rocks below the tillite also have been scratched, or striated, by the movement of the overlying boulder-strewn ice. Another type of glacial deposit is a varved (laminated) sediment composed of alternating millimetre-to-centimetre-thick layers of silt and clay, which closely resemble the layered varves that are laid down in modern glacial lakes at the front of retreating glaciers or ice sheets. Each of these layers defines an annual accumulation of sediment. Varved sediments may contain dropstones, which are fragments of rock that have dropped from an overlying floating ice sheet and that have sunk into and depressed the layers beneath them. When all these features are found together, they provide good evidence of ancient glaciations.

The most extensive early Proterozoic Huronian glaciation occurred 2.3 billion years ago in what is now northern North America. Glacial deposits—similar in age to those of the Huronian—are located in the Transvaal and Cape regions of South Africa, where they reach only 30 metres (100 feet) in thickness but extend over an area of 20,000 square km (7,700 square miles). Such deposits are also encountered in the Hamersley Basin of Western Australia, in east-central Finland and the adjoining part of northwestern Russia, near Lake Baikal in Siberia, and in central India, suggesting the occurrence of a wide-spread glaciation.

Evidence for the largest glaciation in Earth's history, known as the Snowball Earth or Slushball Earth event, dates from the late Proterozoic between 1 billion and 600 million years ago. The principal occurrences of these global glacial deposits are in Europe (Scotland, Ireland, Sweden, Norway, France, the Czech Republic, and Slovakia), the Western Cordillera (Yukon, Can., to California, U.S.) of western North America and the Appalachians of the United States, eastern Greenland, Brazil, much of Africa (Congo [Brazzaville], Angola, Namibia, Zambia, Congo [Kinshasa], and South Africa), and much of Russia, China, and Australia. In addition to the Flinders Range deposits, other notable deposits include the Port Askaig tillite on the island of Islay off northwestern Scotland, which is only 750 metres (2,460 feet) thick but records 17 ice advances and retreats and 27 periglacial periods (which are indicated by infilled polygons that formed under ice-free permafrost conditions). There are two major tillites in central Africa and Namibia (910 to 870 and 720 to 700 million years old, respectively) and two other such consolidated tills in eastern Greenland.

The Correlation of Precambrian Strata

The fact that Phanerozoic sediments have been so successfully subdivided and correlated is attributable to the presence of abundant fossil remains of life-forms that evolved and underwent changes over time. Precambrian sediments lack such fossils, thus preventing any comparable correlations. There are, however, stromatolites in Precambrian sediments ranging in age from about 3.5 billion to 542 million years that reached their peak of development in the Proterozoic. Stromatolites underwent evolutionary changes sufficient for Russian biostratigraphers to use to subdivide the Riphean sequence into four main zones throughout widely separated areas of former Soviet territory. Similar stromatolite-based stratigraphic divisions have been recognized in the Norwegian islands of Spitsbergen, China, and Australia. This stromatolite biostratigraphy still has relatively limited application, however. As a consequence, it is the chronometric time scale that is used to subdivide Precambrian time and to correlate rocks from region to region and from continent to continent.

The rocks within Proterozoic orogenic belts are invariably too deformed to allow correlation of units between different belts. Nonetheless, the techniques of geochronology—in particular, zircon dating—have improved considerably in recent years, with the result that rocks of approximately similar age on different continents can be mutually compared and regarded as equivalent. The isotopic dating of Archean rocks, especially with the use of zircons, has enabled similarities and differences in age to be determined, thereby aiding correlation.

Establishing Precambrian Boundaries

There is no record of tectonic activity of any sort at the time corresponding to the Archean-Proterozoic boundary—about 2.5 billion years ago. This probably means that a supercontinent had been created by the amalgamation of innumerable smaller continental blocks and island arcs. Accordingly, this was a period of tectonic stability that may have been comparable to the Permian-Triassic when the supercontinent of Pangea existed. The main geologic events would have been the intrusion of basic dikes and the formation of sedimentary basins such as the Huronian on the U.S.-Canadian border, into which large volumes of clastic sediment (that is, sediment of predominantly clay, silt, and sand sizes) were deposited. Such sediments would have been derived by erosion of high plateaus and mountains that are characteristic of a large continental mass.

Precambrian Geologic Formations

Examples of Precambrian rocks and fossils occur in several formations throughout the world. Several of the formations described below reference regional rather than global rock systems. In addition, this section contains information on a number of orogenies, or mountain-building events, that occurred during Precambrian time.

Animikie Series

The Animikie Series is the division of Precambrian rocks in North America and the uppermost division of the Huronian System. Rocks of the Animikie Series overlay those of the Cobalt Series.

The Animikie Series was named for exposures along the north shore of Lake Superior, in the Thunder Bay area (*animiki* is the Chippewa word for "thunder"). Animikian rocks occur in Ontario, northern Michigan, northern Wisconsin, and northeastern Minnesota. In the Port Arthur area, Animikian rocks overlie pre-Huronian metamorphic (altered) rocks; rocks of the Cobalt Series are absent. The Animikie Series begins with a basal conglomerate, which is succeeded by very thick sequences of sandstones, shales, and limestones. Animikie rocks are rich in iron ores and provide many important sources for mining. The Mesabi, Gunflint, and Cuyuna ranges are Animikian in age and provide rich deposits of taconite, a mineral assemblage consisting of the iron-containing minerals hematite, siderite, pyrite, and magnetite, and others. Algal structures known as stromatolites, concentric and finely laminated, are common in Animikian limestones. Animikian rocks are commonly intruded by igneous masses but are not strongly folded or faulted.

Avalonian Orogeny

The Avalonian orogeny was a mountain-building event that affected the eastern portion of the Appalachian Geosyncline in late Precambrian time. Evidence for the orogeny consists of igneous intrusions, folding of strata, and the development of angular unconformities in the Avalon Peninsula of Newfoundland, the eastern portion of the Maritime Provinces of Canada, and the southeastern coastal area of New England. Thick sequences of late Precambrian clastic sedimentary rocks in the Blue Ridge and Piedmont provinces of the central and southern Appalachians may also be a reflection of the Avalonian orogeny in these areas.

Banded-Iron Formation (BIF)

BIF is a chemically precipitated sediment, typically thin bedded or laminated, consisting of 15 percent or more iron of sedimentary origin and layers of chert, chalcedony, jasper, or quartz. Such formations occur on all the continents and usually are older than 1.7 billion years. They also are highly metamorphosed. Most BIFs contain iron oxides—hematite with secondary magnetite, goethite, and limonite—and are commonly used as low-grade iron ore (such as the type that occurs in the Lake Superior region of North America). Because BIFs apparently have not formed since Precambrian time, special conditions are thought to have existed at the time of their formation. Considerable controversy exists over BIF origin, and a number of theories have been proposed. Their formation has been variously ascribed to volcanic activity; rhythmic deposition from iron and silica solutions due to seasonal variations; oxidation of iron-rich sediments contemporaneous with deposition; and precipitation from solution as a result of special oxidation-reduction conditions.

Belt Series

This major division of late Precambrian rocks in North America was named for prominent exposures in the Belt Range in southwestern Montana. The thickness of Beltian rocks, which extend northward into Canada, ranges from more than 11,000 metres (about 36,000 feet) on the west to about 4,000 metres (about 13,100 feet) in the east. The upper portions of the Belt Series grade into undoubted Cambrian rocks without apparent interruption, whereas the lower portions are at least 1.5 billion years old, as determined by radiometric-dating techniques.

Four divisions of the Belt Series are recognized. The uppermost, or youngest, of these is the Missoula Group, which is underlain in turn by the Piegan Group and the Ravalli Group; older Beltian rocks are termed Pre-Ravalli. Beltian rocks rest on a basement of gneisses and consist of thick deposits of sandstones, shales, sandy shales, and limestones. Although mud-cracked reddish shales occur, gray shales predominate. Limestones and shales are dominant in the east, whereas to the west sandstones and shales are dominant. It is probable that this coarsening of sediment types to the west indicates the presence in Beltian time of a major landmass still farther to the west, perhaps in the region of what is today the state of Washington or British Columbia. Ripple marks in the gray shales show that most Beltian rocks were deposited in shallow water. The reddish shales were probably deposited on low-lying floodplains, but their origin is disputed.

The remains of Precambrian organisms have been found in the Belt Series and include stromatolites (carbonate-secreting algal mats) and the burrows of wormlike creatures. More advanced fossils attributed to the Beltian are probably of Cambrian age.

Beltian Geosyncline

The Beltian geosyncline is a linear trough in the Earth's crust in which rocks of Precambrian age were deposited in the Northern Rocky Mountain region. The rocks consist of limestones, shales, and sandstones and attain total thicknesses as great as 10,600 metres (35,000 feet). Beltian rocks are exposed in Glacier National Park and in

Precambrian rocks rise above Cretaceous sedimentary material at Chief Mountain, Glacier National Park, Montana. Dr. Marli Miller/Visuals Unlimited/Getty Images the Little Belt Mountains of Montana, from which the name is derived.

Bruce Series

The Bruce Series is a division of Precambrian rocks in North America that is well-developed northeast of the Lake Huron region. The Bruce Series is the lowermost of the three major divisions of the Huronian System. It overlies pre-Huronian schists, underlies rocks of the Cobalt Series, and consists of about 1,500 metres (about 5,000 feet) of quartzites, conglomerates, limestones, and siltstones. The conglomerate that begins the Bruce is separated from pre-Huronian schists by a profound erosional surface representing a time span of unknown duration.

The Bruce sediments were probably deposited by a shallow sea on a locally subsiding, almost flat erosional surface. The upper beds of the Bruce surface are unevenly eroded, and it seems likely that a period of uplift occurred without significant deformation before the deposition of the sediments of the overlying Cobalt Series. The Bruce Series contains significant and valuable deposits of pitchblende and uranite, a source of uranium largely concentrated in the Mississagi Formation, a quartz pebble conglomerate. Controversy surrounds the mode of origin of these ore deposits. Many geologists think that the deposits had a hydrothermal origin. Others feel that the deposits should be classed as sedimentary rocks, and still others think uranium mineralization is related to granitic intrusions in the region. Radiometric dating techniques place the age of the Bruce at about 1.62 billion years.

Canadian Shield

One of the world's largest geologic continental shields, the Canadian Shield is centred on Hudson Bay and extends for 8 million square km (3 million square miles) over eastern, central, and northwestern Canada from the Great Lakes to the Canadian Arctic and into Greenland, with small extensions into northern Minnesota, Wisconsin, Michigan, and New York, U.S.

The Canadian Shield constitutes the largest mass of exposed Precambrian rock on the face of the Earth. The region, as a whole, is composed of ancient crystalline rocks whose complex structure attests to a long history of uplift and depression, mountain building, and erosion. Some of the ancient mountain ranges can still be recognized as a ridge or belt of hills, but the present appearance of the physical landscape of the Canadian Shield is not so much a result of the folding and faulting and compression of the rocks millions of years ago as it is the work of ice in relatively recent geologic time. During the Pleistocene Epoch (2.6 million to 11,700 years ago), the vast continental glaciers that covered northern North America had this region as a centre. The ice, in moving to the south, scraped the land bare of its overlying mantle of weathered rock. Some of this material was deposited on the shield when the ice melted, but the bulk of it was carried southward to be deposited south and southwest of the Canadian Shield.

The resulting surface consists of rocky, ice-smoothed hills with an average relief of 30 metres (100 feet), together with irregular basins, which are mostly filled by lakes or swamps. In places the old mountain ranges may be recognized by hills several hundreds of metres in height. The northeastern portion, however, became tilted up so that, in northern Labrador and Baffin Island, the land rises to more than 1,500 metres (5,000 feet) above sea level.

Coutchiching Series

A division of rocks in the region of northern Minnesota and Ontario, the Coutchiching Series has been dated radiometrically to about 2.6 billion years ago. Rocks of the

Coutchiching Series appear to underlie those of the Keewatin Series, at least in some areas, and consist of mostly sedimentary rocks that have been altered to varying degrees by metamorphic processes. Some geologists consider the Coutchiching older than the Keewatin Series, but others dispute this view. Much study on the stratigraphic relationships between Coutchiching and Keewatin strata remains to be done before a conclusive answer can be provided. It seems likely that much of the Coutchiching Series is interbedded with lava flows assigned to the Keewatin Series and that the two are at least partly equivalent in age.

Dalradian Series

The Dalradian Series is a sequence of highly folded and metamorphosed sedimentary and volcanic rocks of late Precambrian to Early Cambrian age, about 540 million years old, that occurs in the southeastern portions of the Scottish Highlands of Great Britain, where it occupies a belt 720 kilometres (450 miles) long.

Containing no fossils over most of its outcrop area, the Dalradian has yielded rare specimens of the trilobite genus *Pagetides,* a Lower Cambrian form known from North America, collected from strata near the top of the series. Thus, it is known that the upper portion of the Dalradian is Lower Cambrian, but the position of the Precambrian-Cambrian boundary is uncertain.

Metamorphism—chemical and physical alteration from elevated temperatures and mechanical stresses in the Earth's crust—related to the Caledonian orogenic (mountain-building) episode, has not obscured the original nature of Dalradian sedimentary types. Primary features are still evident in quartzites, calcareous limes, conglomerates, and graywackes, whereas fine-grained sedimentary rocks are represented by slates, phyllites, and schists. Current bedding can be seen in the graywackes and quartzites. Combined thicknesses of the Dalradian are thought to approach 9,100 metres (30,000 feet) and probably represent geosynclinal (downward flexures in the Earth's crust) accumulations.

Dalradian sequences are also known from Ireland, especially in northern Donegal.

Grand Canyon Series

The Grand Canyon Series is a major division of rocks in northern Arizona that dates from Precambrian time. The rocks of the Grand Canyon Series consist of about 3,400 metres (about 10,600 feet) of quartz sandstones, shales, and thick sequences of carbonate rocks. Spectacular exposures of these rocks occur in the Grand Canyon of the Colorado River in northwestern Arizona, where they overlie the strongly deformed and contorted Vishnu Schist—the angularity of which stands in bold contrast to the almost horizontal bedding of the Grand Canyon Series. The Grand Canyon Series actually dips slightly eastward and is separated from the overlying Cambrian sandstones by a major erosion surface unconformity. A conglomerate was deposited on the eroded surface of the Vishnu Schist. Limestones, shales, and sandstones occur over the conglomerate and are thought to represent shallow water deposits. The area of deposition was probably a large deltaic region that was slowly subsiding, allowing great thicknesses of sediment to accumulate near sea level. The presence of Precambrian organisms is indicated by calcareous algaelike structures in the carbonate rocks, as well as by tracks and trails of wormlike creatures in other rocks. Initially, in a generalized outline of the Precambrian history of the region, the Vishnu Schist was upraised, folded, and metamorphosed and then slowly eroded and

Layers of Precambrian rock can be seen in the Grand Canyon, northwestern Arizona. Shutterstock.com

worn down to a flat surface. The Grand Canyon Series was deposited perhaps as part of a slowly subsiding geosynclinal trough. The region was then subjected to uplift and tilting, and a Precambrian period of erosion for the Grand Canyon Series began. This action was later followed by a long period of deposition during the Paleozoic Era (542 to 251 million years ago) and then further erosion during the Cenozoic Era (beginning 65.5 million years ago) until the region assumed its modern configuration.

Hudsonian Orogeny

The Hudsonian orogeny was a Precambrian thermal event on the Canadian Shield that took place 1.7 billion years ago (± 1.5 million years). Rocks that produce dates in this time span are those in the Churchill Province, a large arcuate belt that includes most of Canada west of Hudson Bay, the exposed Precambrian rocks in northern Canada, the Arctic Islands and Baffin Land, most of northern Greenland, and part of Labrador; the Bear Province in the northwestern tip of the shield; and a broad area in the western United States. The Mazatzal orogeny in Arizona, the Black orogeny in South Dakota, and the Penokean orogeny in the southern part of the Lake Superior region may represent the Hudsonian event in the United States. Precambrian rocks in the Southern Province, which extends south-southwest of Lake Superior into the mid-continental United States, also are dated in the Hudsonian time span.

Huronian System

A major division of Precambrian rocks in North America, the Huronian System is well known in the Great Lakes region and has been divided into three major series of rocks: the lowermost, the Bruce Series, is followed in turn by the Cobalt and Animikie series. The Huronian System forms a wide belt of sedimentary rock units along the north shore of Lake Huron and consists of about 4,000 metres (about 12,000 feet) of sandstones, shales, and conglomerates. The sequence is more complete and thicker to the west, where thicknesses of about 6,000 metres (20,000 feet) of Huronian rocks occur. Important iron-bearing Huronian rock units found in northern Wisconsin and central Minnesota are of major economic significance.

Indian Platform

The Indian Platform is another Precambrian continental shield, one of four around which the Asian continent coalesced. Five areas of geosynclinal folding constitute the platform basement. From oldest to youngest, these are the Dharwar, Aravalli, Eastern Ghat, Satpura, and Delhi foldings. After the consolidation of the basement in Proterozoic times (between 2.5 billion and 542 million years ago), the evolution of the platform comprises six major tectonic-stratigraphic sequences—that which produced the rocks of the Cuddapah Group and those of the Vindhyan, Gondwana, Mesozoic, Paleogene, and Neogene sequences.

Katangan Complex

The Katangan Complex is a major division of late Precambrian rocks in central Africa, especially in Katanga province, Congo (Kinshasa). The Katangan Complex is a complicated array of diverse sedimentary and metamorphic rocks. Katangan rocks consist of shales, quartzites, limestones, sandstones, dolomites, and slates more than 7,000 metres (23,000 feet) thick. The Katangan has been radiometrically dated at more than 620 million years in age. Katangan rocks are of tremendous economic importance; copper, cobalt, uranium, zinc, and other valuable minerals are abundant.

Kenoran Orogeny

The Kenoran orogeny was a Precambrian thermal event on the Canadian Shield that occurred 2.5 billion years ago (± 150 million years). Rocks affected by the Kenoran event represent some of the oldest rocks in North America and occur in the Superior Province surrounding Hudson Bay on the south and east, the Slave Province in northwestern Canada, and the small Eastern Nain Province on the northeastern Labrador coast.

Beyond the shield area, rocks equivalent in age to Kenoran rocks occur in Wyoming and the Black Hills. Parts of the Lewisian Gneiss of Scotland have been dated at 2.4 billion to 2.6 billion years ago, and these have been correlated with the Kenoran thermal event.

Keweenawan System

The Keweenawan System is a division of late Precambrian rocks and time in North America. Rocks of the Keweenawan System are about 10,700 metres (about 35,000 feet) thick, overlie rocks of the Huronian System, and underlie rocks of the Cambrian System. It has been suggested that the youngest Keweenawan rocks actually may be Cambrian in age. In the Lake Superior region, Keweenawan rocks consist of reddish sandstones, siltstones, shales, and some conglomerates. Great thicknesses of lava flows also occur; it has been estimated that about 100,000 cubic kilometres (24,000 cubic miles) of lava were produced. The burden of the great weight of lava caused the crust beneath to sag and produced the basin that Lake Superior now occupies. The Keweenawan System has been divided into Lower, Middle, and Upper series. The lavas are primarily concentrated in the Middle Keweenawan Series, whereas the Lower Keweenawan Series is dominated by sediments. The Keweenawan System is named for prominent exposures studied at Keweenaw Point, Michigan.

Lewisian Complex

The Lewisian Complex, which is also known as the Lewisian Gneiss, is a major division of Precambrian rocks in northwestern Scotland. In the region where they occur, Lewisian rocks form the basement, or lowermost, rocks. They form all of the Outer Hebrides, as well as the islands of Coll and Tiree, and are exposed along the northwestern coast of Scotland. The oldest rocks of the Lewisian have been dated by radiometric techniques at between 2.4 billion and 2.6 billion years old, whereas the youngest Lewisian rocks have been dated at 1.6 billion years. Lewisian rocks originally consisted of both igneous and sedimentary rocks that have been altered from their original composition and structure through time by heat, pressure, and the action of solutions of one sort or another. The dominant rock type is grayish gneiss that is rich in quartz, feldspar, and iron-rich minerals. Some sedimentary-derived Lewisian rocks, especially in the region of Loch Maree and South Harris, still retain some of their original sedimentary features and indicate that the sediments were originally shales, sandstones, and some limestones. Many igneous intrusions also occur, including granites, pegmatites, and dolerites. Three major subdivisions of the Lewisian Complex are recognized: the lowermost Scourian Complex, followed by the Inverian Complex and the Laxfordian Complex. Rocks of the Lewisian Complex are overlain by those of the Torridonian Series. Lewisian rocks have been profoundly affected by two major periods of deformation, the first of which occurred during the time represented by the Scourian Complex and the second during the Laxfordian. The radiometric dates obtained for the age of the Lewisian are essentially the dates of these periods of deformation.

Longmyndian Series

This major division of Late Precambrian rocks and time occurs in the southern Shropshire region of England. Named for prominent exposures in the Longmynd Plateau region, Longmyndian rocks consist of steeply angled and even overturned unfossiliferous mudstones, sandstones, conglomerates, and volcanic rocks. Two major subdivisions are recognized: the Western Longmyndian and the underlying Eastern Longmyndian. The Western Longmyndian consists of the Wentnor Series, purple sandstones, conglomerates, and some greenish siltstones and shales. Thicknesses of about 4,800 metres (15,700 feet) of Wentnor rocks have been measured. The Eastern

Longmyndian is subdivided into the overlying Minton Series and the underlying Stretton Series. The Minton Series, about 1,200 metres (about 3,900 feet) in thickness and made up of purple and green shales, sandstones, and conglomerates, is separated from the underlying Stretton Series by an unconformity representing a period of erosion rather than deposition. The Stretton Series, grayish and greenish siltstones, sandstones, shales, and volcanic rocks, is as much as 3,500 metres (about 11,500 feet) thick. Rocks underlying the Stretton Series and possibly related to the Longmyndian are known as the Eastern and Western Uriconian, geographically separated from each other but similar in lithology and probably broadly contemporaneous. The Eastern and Western Uriconian consist of lavas, tuffs, and intrusive igneous bodies. They are separated from the overlying Stretton Series by a prominent unconformity. Elsewhere, in the Charnwood Forest and Midlands regions, a sequence of rocks occurs that may favourably be compared to the Stretton Series of the Eastern Longmyndian. Three subdivisions have been recognized: the lowermost Blackbrook Series, overlain in turn by the Maplewell Series and the Brand Series. These rocks, collectively known as the Charnian, consist largely of volcanic rocks (most prominent in the Maplewell Series and least in the Brand Series) and of sedimentary conglomerates, sandstones, siltstones, and slates.

Charnian sedimentary rocks contain impressions of a Precambrian organism known as *Charnia*; these are especially prominent in the higher levels of the Maplewell Series. Similar if not identical forms are known to occur in Australia. The zoological affinities of *Charnia* are uncertain. Opinions have ranged from including the form in the Coelenterata (corals, hydras, and jellyfish) to the algae.

Onverwacht Series

A division of Archean rocks in the Swaziland region of southern Africa, the Onverwacht Series is well known from exposures in the Komati valley in the eastern Transvaal region, South Africa. Onverwacht rocks consist of dark, andesitic lavas, dolomitic limestones, cherts, and jaspers, as well as serpentines, gabbros, and peridotites. Many of the rocks have been intensely deformed by metamorphic processes. The results of these processes are observed as locally abundant schists and marbles. Rocks of the Onverwacht Series underlie those of the Fig Tree Series and form the basement rocks in the region of their occurrence.

Radiometric dating techniques have established the age of the Onverwacht Series at about 3.7 billion years. Tiny cell-like forms have been discovered and studied in the Onverwacht rocks of sedimentary origin; studies of the hydrocarbon compounds in the rocks have also been carried out. The Onverwacht Series may contain some of the earliest known traces of living organisms on Earth. Indeed, it is possible that the transition from nonliving to living material is recorded within the Onverwacht. Further study, it is hoped, will elucidate the actual steps in this transition.

Pound Quartzite

The Pound Quartzite is a formation of Precambrian rocks in the region of Adelaide, South Australia. The formation consists of shales and siltstones, limestones, and quartzites. It is notable because, from it, a very early fossil assemblage, the Ediacara fauna, was recovered. The fossil assemblage evidences a variety of types, many of which are surprisingly complex and afford a rare glimpse at the diversity of Precambrian animal life.

Seine Series

The Seine Series is a division of Precambrian rocks that occur in Ontario and northern Minnesota named for prominent exposures studied along the Seine River, Ontario. It forms a thick sequence of sedimentary rocks that overlie the Keewatin Series and are separated from it by an unconformity, a surface representing a major period of erosion before sediments of the Seine Series were deposited. The series begins with a prominent conglomerate that includes large boulders of granite within its matrix. It is thought that this conglomerate represents a long period of erosion during which the granitic cores of Keewatin mountains were exposed. Granitic intrusions occur in the Seine Series and have been dated at about 1.1 billion years old, and the granites that formed the mountain cores have been dated at about twice that. The Seine Series may be equivalent to the Knife Lake Series. If so, the term Knife Lake Series would have priority over Seine Series.

Sturtian Series

This division of Proterozoic rocks found in south central Australia forms the lower part of the Umberatana Group, a large unit of rocks made up of siltstone, silty shale, quartzite, limestone, and conglomerate. The Sturtian Series is partly interpreted as being of glacial origin from the glacially produced pavements that have been recognized. The Sturtian begins with a boulder-laden horizon that generally overlies quartzites. Tillites, cemented glacial till deposits, occur higher up in the Sturtian sequence and include many glacial erratics seen as striated and faceted boulders of granite, gneiss, quartzite, shale, and limestone. Silty shales that may be laminated or even exhibit glacial varves also occur. The Sturtian thickens to the north and northeast of Adelaide, where it may be as much as 6,000 metres (20,000 feet) thick. Above the glacial deposits of the Sturtian lie blue-gray shales, arkoses, shaly sandstones, siltstones, and dolomites.

The Sturtian Series is the most widespread unit of the Precambrian sedimentary basin in south central Australia, the Adelaide Geosyncline. In the Olary region, the Sturtian is intruded by granite igneous masses, whereas in the Everard Ranges basalt flows are found with the tillites. The Sturtian Series is overlain by the Marinoan Series, which similarly includes within it a sequence of glacial deposits.

Swaziland System

The Swaziland System is a major division of rocks and time in southern Africa in Precambrian time. The system consists of a great thickness of sedimentary and metamorphic (altered) rocks with numerous intrusions of igneous bodies. Two major subdivisions of the Swaziland System are recognized, an Upper and a Lower series. Many of the units that constitute the Swaziland System contain mineral deposits of great value, including gold, copper, and uranium—a circumstance that has spurred intensive study and exploitation.

Waterberg Series

The Waterberg Series, which is also called the Waterberg System, is a major division of rocks in southern Africa. The age of the Waterberg is in doubt; it is possible that the Waterberg is late Precambrian or Early Paleozoic (older or younger than 542 million years, respectively). Waterberg rocks consist of several thousand feet of brown, red, and purple sandstones and some rather minor

The Waterberg Plateau of Namibia, southern Africa, is made up of Precambrian rock. © www.istockphoto.com/brytta

shales. Sedimentary structures such as ripple marks, desiccation cracks, and current bedding are prominent features in the almost flat-lying Waterberg rocks. As yet no evidence of fossilized organisms has been found in them.

Witwatersrand System

This major division of Precambrian rocks in South Africa overlies rocks of the Dominion Reef System, underlies those of the Ventersdorp System, and occurs in an east-west band from Randfontein to Springs and from the Vaal River in the region of Klerksdorp in the north to Ventersdorp in the south. The rocks actually occupy a much larger area. Much of the Witwatersrand System is covered by later deposits, and the subsurface areal extent of Witwatersrand rocks has been delimited by exploratory geophysical and drilling studies because the Witwatersrand is of great economic importance owing to its valuable deposits of gold and uranium.

In all, the Witwatersrand System consists of about 8,100 metres (26,600 feet) of rocks that have been segregated into an upper and lower division, each of which is further divided into series. Three series are recognized in the lower division: the lowermost Hospital Hill Series, the Government Reef Series, and the Jeppestown Series, respectively. The upper division is divided into the lower Main-Bird Series, followed by the Kimberley-Elsburg Series. The Government Reef Series consists of alternating shales and quartzites in addition to pebbly layers that contain gold deposits. It also contains indications of a period of extensive glaciation. The most economically important series is the Main-Bird Series, largely quartzitic conglomerates that are extremely rich in uranium and gold. Large quantities of gold are also found in the Kimberley-Elsburg Series of shales, quartzites, and dolomites.

CONCLUSION

Establishing the definitive timeline for Earth's history is an ongoing process that has spanned several generations of thinkers. Early scientists, such as Nicolaus Steno and Johann Gottlob Lehmann, were the first to notice the sequential nature of the layers of rock they examined. The significant contributions of James Hutton and Charles Lyell advanced the idea of gradual change in rocks, while Georges Cuvier did the same with fossils. Other thinkers, such as William Smith and Giovanni Arduino, were the first to correlate rocks. The first concerted attempts of establishing an order to geologic time did not occur until the 19th century with the work of Adam Sedgwick, Roderick Impey Murchison, Charles Lapworth, and others.

Since then, technological advances—such as mass spectrometry and dating techniques based on the decay of radioactive elements—have allowed the efficient correlation rocks from different continents. This work has enabled geologists, paleontologists, and other scientists to construct a timeline of Earth's history from the planet's formation about 4.6 billion years ago to the present.

The first step in the journey through geologic time began with the Precambrian, a long interval characterized by Earth's violent creation, the evolution of the atmosphere and oceans, the formation of continents, and the emergence of life. In later intervals, as continents continued to break apart and come together, mountain ranges rose while coastal plains sank, Earth's climate oscillated between warmth and cold, and life spread throughout the seas and land.

Glossary

alluvial A type of sedimentary material deposited by rivers, consisting of silt, sand, clay, gravel, and organic matter.

banded-iron formation Chemically precipitated sediment, typically thin bedded or laminated, consisting of 15 percent or more iron of sedimentary origin and layers of chert, chalcedony, jasper, or quartz.

basalt Extrusive igneous (volcanic) rock that is low in silica content, dark in colour, and comparatively rich in iron and magnesium.

craton The stable interior portion of a continent characteristically composed of ancient crystalline basement rock.

dendrochronology Also called tree-ring dating, the field of science that uses growth rings in trees to date and interpret past events, particularly paleoclimates and climatic trends, based on the analysis of tree rings.

discoidal Of, resembling, or producing a disk.

igneous rock Any of various crystalline or glassy rocks formed by the cooling and solidification of molten earth material.

leaching Loss of soluble substances and colloids from the top layer of soil by percolating precipitation.

metamorphic rock Any of a class of rocks that result from the alteration of preexisting rocks in response to changing environmental conditions, such as variations in temperature, pressure, and mechanical stress, and the addition or subtraction of chemical components.

Neptunist theory Scientific theory positing that all the world's rocks were formed by sedimentation from the oceans.

paleontology Scientific study of life of the geologic past that involves the analysis of plant and animal fossils, including those of microscopic size, preserved in rocks.

paleosols Fossilized soil deposits preserved by burial underneath either sediments or volcanic deposits, which in the case of older deposits have lithified into rock.

petrifaction The process by which organic material is converted into stone by impregnation with silica.

photosynthesis The process by which green plants and certain other organisms transform light energy into chemical energy.

primordial Existing in or persisting from the formation of the Earth.

protozoan Any member of the subkingdom Protozoa; a collection of single-celled eukaryotic organisms.

red beds Continental sediments dominated by sandstones and shales of a red colour.

sedimentary rock Rock formed at or near the Earth's surface by the accumulation and lithification of sediment or by the precipitation from solution at normal surface temperatures.

strata Sedimentary rock layers bounded by two stratification planes, the latter being produced by visible changes in the grain size, texture, or other diagnostic features of the rocks above and below the plane.

terrestrial Relating to the land as distinct from air or water.

till Unsorted material deposited directly by glacial ice and showing no stratification.

tillites Sedimentary rock that consists of consolidated masses of unweathered blocks and glacial till in a rock flour (matrix or paste of unweathered rock).

uniformitarianism The paradigm that posits that existing processes that act upon Earth in the present also existed in the past.

volcanism Any of various processes and phenomena associated with the surficial discharge of molten rock, pyroclastic fragments, or hot water and steam, including volcanoes, geysers, and fumaroles.

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THE PALEOZOIC ERA: DIVERSIFICATION OF PLANT AND ANIMAL LIFE

THE MESOZOIC ERA: AGE OF DINOSAURS

THE CENOZOIC ERA: AGE OF MAMMALS

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