地 质 专 业 英 语

（阅读与写作）

Reading and Writing Materials for Geological English

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第一部分　地质专业文献阅读

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How to use a rock section to tell a story: the grand canyon

There are many places where the bedrock, which everywhere underlies the surface, is exposed, not obscured by soil or loose boulders. Such exposures, called **outcrops**, range in size from small projections of weathered rock on a hillside and ledges in the beds of small streams to high cliffs that make the walls of canyons in mountainous terrain. Geologists also know how to use such engineering works as road cuts to study well-exposed sections of rock. No highway cut, however, could touch the dimensions of a place like the Grand Canyon of the Colorado River. And no cut has the beauty and grandeur of that fantastic gorge, which is more than a mile (1.6 kilometers) deep in places, 4 to 18 miles wide (6.5 to 29 kilometers), and 280 miles (450 kilometers) long. The main part alone is 56 miles (90 kilometers) long. It is still an adventure to travel down the Canyon of the Colorado River in a small boat, repeating the first trip made in 1869 by geologist Major John W. Powell, who later helped found the U.S. Geological Survey. At every turn in the Canyon, from steep rims to inner gorge, one sees rock in a multitude of forms large and small. Revealed are patterns of rock characteristics that we use to reconstruct geologic history.

The first thing we notice about the rocks of the Grand Canyon is a pronounced horizontal **layering**, or **stratification**. This **bedding**, as it is most frequently called, is what we might expect from the settling of particles from air or water to form layers of **sediment**, (from the Latin sedimentum, settled). This expectation is supported by experience, for we can see horizontal layers of sand being deposited on beaches or sandbars and layers of mud and silt accumulating on flood plains of rivers.

Once we make this generalization (which has held up for the three hundred years since it was first enunciated), we can go on to the obvious next thought. Since it is absurd to think that a sedimentary layer can be deposited beneath a previously deposited bed, we conclude that any new layer added to a series must always be added on top. Naturally we have to add the proviso that the whole series has not been deformed and completely overturned at some later time. The time sequence of stratification is the simple basis for the **stratigraphic time scale**, a clock we can use to measure time and date events.

Simple as these generalizations are, they are a beautiful example of a well-known dictum: the truly great discoveries are the ones that are perfectly obvious after someone has pointed them out to us. It was Nicolaus Steno, a Danish court physician living in Italy, who in 1669 formulated the **principle of original horizontality** and the **principle of superposition**, which we have given above in modified form. He also stated the principle of original continuity, which holds that a sedimentary layer forms at the time of its deposition a continuous sheet that ends only by thinning to disappearance, by gradually changing to a bed of different composition, or by abutting against a wall or barrier such as a shoreline that confines the depositional area. From the law of continuity we intuitively grasp the idea that the face of a bed, as we see it in the excavation for a highway cut or in the walls of the Grand Canyon, is the broken or eroded edge of a once- continuous sheet.

From these three principles we get the rudiments of our stratigraphic clock; that is, we establish a total length of time necessary for all of the rocks to be laid down and a time interval within the whole span for each layer. If we had some idea of how long each bed took to be laid down, and **if** all of the time span were accounted for by the time to lay down the sum of all of the beds, our clock would be constructed.[1]

Unfortunately, that last **if** is a big one. Based upon the observation of river floods and other kinds of sedimentation, we suspect that a certain amount of time is not represented by rock. The silts laid down on floodplains of rivers, such as the historic ones of the Nile River in ancient Egypt, do not accumulate steadily and uniformly. The time scale of flood deposits is about days long, but there is also a time scale for the times between floods, a time interval that may range from a few years to several decades. In other words, a hiatus, or interruption in sedimentation, may be two to three orders of magnitude (powers of 10) greater than the time for deposition of a layer of flood silt.

How old is a fossil?

We have another, more powerful tool for establishing the time sequence of a series of layers of sedimentary rocks, and that tool is fossils, the remains of ancient organisms found in some of those rocks. **Limestones,** one class of sedimentary rock, are made up of calcium carbonate ($CaCO_3$), much of it in the form of fragments of fossil animal shells. **Shales,** rocks that are hardened and compacted clays and muds, and **sandstones**, rocks made of cemented sand grains, may also contain fossil materials such as shells and shell fragments. Some of the fossil shells found in such rocks can easily be identified by comparison with similar shells found today. Many others look vaguely like some animals that live today but are obviously different, and others are obviously some sort of organism’s shell but not like anything alive today. Not all fossils are shells of invertebrates like clams, oysters, cowry shells, and periwinkles. Diggers in river beds millions of years old may uncover bones of vertebrates such as reptiles or mammals—sometimes even dinosaur remains. Fish skeletons and shark teeth are found in other rocks. Plant fossils are abundant in some rocks, particularly the rocks in and above coal beds, where one can recognize fernlike and other leaves, twigs, branches, and even whole tree trunks.

What could be more natural than to jump to the conclusion that these fossils represent life of former times and that from them we could deduce the flow of evolution from the most primitive organisms to such a complicated one as modern *Homo sapiens.*

[1] A special kind of sediment used to tell time, varved clay, forms in lakes that freeze over in the winter. The varves are layers of the clay that form couplets; one part of each couplet is a relatively thick, coarse-grained, light-gray silty clay that grades upward into a thin, fine-grained dark-gray clay. The light layers form in the summer and the dark ones in the winter, each pair marking one year. By counting the annual banding in lakes formed after the last ice age in northern Europe, Baron G.DeGeer of Sweden was able to determine that about 8,700 years had passed since the glacial ice had retreated from all of southern Europe. The accuracy of this method was confirmed by radioactive age determinations made much later, and only slight revision of the date had to be made.

The “jump” took some time to make, though. One of the first to make it in modern times (some Greeks had known it long before) was Leonardo da Vinci. He was followed later by Nicolaus Steno, who in the seventeenth century compared teeth from modern sharks with the so–called tongue-stones of Malta in the Mediterranean and concluded that they must have come from the same kind of shark. Many objected to Steno’s conclusion, but the similarities between the forms of modern animals, especially the hard parts, such as teeth, bones, and shells, soon piled up in such numbers that the evidence was overwhelming and could not reasonably be dismissed either as “some accident of form” or as an obscure expression of God’s wisdom in creating the Earth.

But what has all this to do with making a time scale? The beginning of an answer can be found in the rocks of the Grand Canyon, for there we can find fossils in many of the exposed rocks, particularly in the limestones. Each laver of limestone has a number of fossils of different species, but the assortment of fossils in one limestone may differ sufficiently from those in another as to be readily distinguishable. This vertical arrangement of different fossils is called the **faunal succession**. The sequence of life forms the fossils represent corresponds to the series of sedimentary rock layers bearing the fossils, the **stratigraphic sequence**, and the faunal and stratigraphic series have the same order. For convenience in mapping, limestones and other rocks are grouped into **formations**, groupings of layers that are everywhere the same stratigraphic age and contain materials that for the most part have the same physical appearance and properties. This combination of appearance and properties is called **lithology.** Formations are a convenience for mapping. Some consist of a single rock type or lithology, such as limestone. Others may consist of interlayered thin beds of different lithologies, such as sandstone or shale. However they vary, each formation comprises a distinctive set of rock layers that can be recognized and thus mapped as a unit.

Once formations and stratigraphic sequences all over the world had been mapped by nineteenth-century geologists, it became apparent that everywhere the faunal successions matched the sequences. This is the rule in fossiliferous formations of all ages since the beginning of the Cambrian Period, when shelled animals evolved[2].

Thus fossil assemblages can be used as “fingerprints” of formations; each assemblage has distinguishable characteristics, even though individual species may be present in several different formations.

It was this feature of fossil-bearing sediments that William Smith, an engineer and surveyor who had worked in coal mines and mapped along canals, began to see so clearly when he set about collecting fossils in southeastern England in 1793. Smith knew nothing of the idea of organic evolution that Charles Darwin was to enunciate some decades later. He did note, however, that different formations contain different fossils, and he was able to tell one formation from another by the differences in the fossils. As Smith extended his mapping over much of southern England in the early part of the nineteenth century, he was able to draw up a stratigraphic succession of rocks that

[2] Up to the mid-twentieth century, geologists believed there were no fossils in rocks older than the Cambrian—called the Precambrian. We now know of the remains of single-celled organisms throughout most of Precambrian times. Paleontologists and paleobotanists are working now to erect faunal successions for these early times.

appeared in different places at different levels—a composite showing how the complete section would have looked if all of the rocks from those different places and levels were brought together in a single place.

The grand canyon sequence interpreted

Let us return to the Grand Canyon and work through the sequence from the base up. At the bottom, exposed in the inner gorge, are dark rocks that are not at all like the horizontally bedded sedimentary rocks above. They show no bedding and form bodies that seem to be inserted into and cut across structures in the surrounding rocks. Some of these rocks are made up of coarse crystals; others, of particles so small that they cannot be seen even with a hand lens. These characteristics are interpreted by geologists as evidence of **igneous** origin; that is, these rocks were formed by cooling and solidification of a hot molten liquid, or **magma**. The coarsely crystalline ones are inferred to have been emplaced in the rocks that surround them while those rocks were still buried deep in the earth; these **intrusives** originated as hot magma that pushed its way into cracks and other openings in the surrounding rocks. The large crystals are characteristic of intrusives and the result of the slow cooling of magma that takes place far below the surface. The fine-grained rocks, **extrusives**, were formed as lava flows and ash deposits from volcanic eruptions. Their characteristic fine texture indicates rapid cooling at the surface.

Another group of rocks exposed in the inner gorge are those that have a platy or leafy texture, called **foliation**, caused by the alignment of minerals along straight or wavy planes. Foliation can be confused with bedding. In some of these foliated rocks true bedding can also be seen, but it is folded and crumpled. These **metamorphic** rocks (from the Greek, meaning to change form or to transform) were once sedimentary and igneous rocks, but they have been altered by the action of heat and pressure due to their deep burial in the Earth.

The lowest rocks of the inner gorge, the Vishnu formation, are a complex mixture of these igneous and metamorphic rocks. Such rocks have no fossils, and there is no way to tell how old a rock is merely by looking at its minerals and their texture (geologists learned long ago that rocks formed by the same process at different times tend to look alike). Nevertheless, the Vishnu is at least the oldest rock we can see in the Grand Canyon inner gorge: it’s at the bottom. Because of its position, it gives us our first glimpse of history there. The rocks of the Vishnu, originally formed as lava flows, ash deposits, and sediments, were deeply buried by the rocks that now overlie them, were metamorphosed by the resulting heat and pressure, and were later intruded by molten magma.

Above the Vishnu, and separated from it by a sharp line of discontinuity—an **angular unconformity**—is another series of rocks. An angular unconformity is a surface of erosion that separates two sets of beds whose bedding planes are not parallel. It thus signifies that the originally horizontal lower set was deformed and then eroded to a more-or-less even surface before the upper set was deposited horizontally upon it. Rocks in many places show evidence of such physical deformation. Sedimentary layers, once horizontal, are in places **folded** (bent into a wavy structure) and **faulted** (broken and displaced along fractures). The same structural features, sometimes less easily recognized, are found in igneous and metamorphic rocks. We can therefore add an episode of deformation and uplift to our history of sedimentation, burial, and metamorphism.

The rocks above the Vishnu are interlayered sandstones, shales, and limestones. The Grand Canyon Series, as these rocks of the inner gorge are called, contains no fossils of shelled organisms such as those of Cambrian and younger age rocks and therefore cannot be tied to a standard faunal succession. All that we can say about the age of this Series from this inspection is that it is younger than the Vishnu, older than the rocks above, and is tilted from its originally horizontal position. Sedimentary rocks like these are perfectly ordinary in every way except for two characteristics. They contain no shelled fossils—though they may have remains of microorganisms such as algae—and they may be associated with deformed and metamorphosed rocks like the Vishnu. In the nineteenth century rocks of this kind were set apart from younger fossiliferous strata and called the Precambrian. Though they were then thought always to be complexly folded and faulted as contrasted with much less deformed younger rocks, we now know many places where this distinction is invalid. From the rocks of the Grand Canyon Series we have a glimpse of a history after the erosion of an uplifted Vishnu terrain that includes sedimentation, moderate burial, and moderate deformation that led to tilting and gentle folding.

Another clearly discernible unconformity separates the Grand Canyon Series from the overlying pebbly brown Tapeats Sandstone. The Tapeats contains no fossils, but it can be dated by reference to the overlying formations because it blends into them without any break, thus forming a **conformable succession**. Farther up the canyon wall, the Tapeats gradually gives way to a formation consisting mainly of shale, a hardened equivalent of a muddy sediment, called the Bright Angel Shale.

The Bright Angel Shale contains a few fossils, most of which are **trilobites**, extinct arthopods related to modern crayfish. The differences among the trilobites of different ages can be used by paleontologists to date these rocks. By matching fossil trilobite species in different stratigraphic sequences in different parts of the world, a composite succession has been worked out. As a result geologists learned that the part of the Bright Angel Shale lying just above the nonfossiliferous Tapeats Sandstone in the western part of the Canyon is older than the part of the Shale that occupies the same stratigraphic position in the eastern part of the Canyon. This means that the sea in which the Bright Angel was deposited flooded the land in the east at a later date. This is evidence of a **transgression**, meaning that as the Bright Angel Shale was being deposited the sea gradually moved landward. Once again, simple geometric evidence leads to this conclusion: as the sea advanced slowly from west to east, it continuously laid down sand along beaches and mud in deeper water. The reverse, the withdrawal of a sea and the inverse distribution of sediments in relation to shorelines, is called a **regression**. The Bright Angel Shale grades upward into the overlying Muav Limestone, also a part of the transgressive sequence.

Even from a great distance, most of the formations of the canyon wall can be distinguished easily. For example, the Tapeats Sandstone forms an obvious rim at the top edge of the inner gorge. In contrast, most of the Bright Angel Shale, just above the rim, is hidden beneath rubble, which slopes gently upward from the top of the Tapeats to the base of a higher step, the base of the Muav Limestone.

The next formation up, the Temple Butte Limestone, could easily be missed. It is thin, and in places along the Canyon wall it is missing entirely; in these places, the Redwall Limestone, which elsewhere overlies the Temple Butte, rests directly on the Muav. The importance of the Temple Butte is that it contains fossil skeletons of primitive fish. We know from the general succession of fossil animals that these fish lived at a much later time than the trilobites of the Muav. Fossils of many marine animals that lived between the times of deposition of the Muav and the Temple Butte are known from other formations in various parts of the world—but they are all missing here. Thus there is evidence here of a great gap in the record, an unconformity between the Muav and the Temple Butte. If any sediment had been laid down during the time represented by this unconformity, it was later eroded without leaving a trace. The sequence implies a history of Muav sedimentation and burial (but no deformation; it is still horizontal) before it was uplifted, eroded, and later covered by sediments that formed the Temple Butte Formation.

An unconformity between the Temple Butte and the overlying Redwall Limestone[3] represents another time gap, and an unconformity between the Redwall and the Supai Formation, yet another. The Redwall's age we know from its sparse content of fossil marine animals. The Supai contains no marine fossils, but it does contain fossils of land plants like those found in the coal beds of the United States and the Ruhr Valley in Europe. Thus the Supai can be dated with reference to the worldwide succession of plant fossils, a succession that has been worked out in much the same way as the succession of animal fossils. Of even greater interest are the fossil footprints of primitive reptiles found in the Supai. Thus it turns out that we can tell something of time not only from marine sediments but from those deposited on land—the **terrestrial**, or **continental, deposits**.

Continuing up the canyon walls, we find another unconformity at the top of the Supai. Above the Supai is a sandy red shale, the Hermit Shale, which is succeeded by the Coconino Sandstone. Not only does the Coconino contain more vertebrate animal tracks, but it has a distinctive form of bedding, which is not uniform and horizontal like that of many sediments but is composed of many sets of interfering wedges of bedded material inclined at angles up to 35° from the horizontal. This form of bedding, called **cross-bedding,** is characteristic of sand dunes on land and of dunes formed by currents in rivers and under the sea. On the basis of vertebrate animal tracks and the dune type of cross-bedding, most geologists believe that the Coconino Sandstone formed by wind.

The Coconino appears to be conformable with the underlying Hermit and, with the overlying Toroweap, a formation of limestone and red, sandy shale. Next upward is a massive formation of limestone and sandstone that forms the top of the cliffs at the upper rim of the Canyon—the Kaibab Formation.

[3] How the Redwall Limestone got its name is obvious from any look at it in the Canyon. But the "red" in Redwall is undeserved in at least one way, for when a piece of the Redwall is cracked the fresh rock turns out to be gray. A close look at the cliff shows that the limestone is stained, as if it had been rubbed with a reddish pastel chalk. The stain is washed down by rain from the overlying sandy red shale of the Supai Formation.

If we were to inspect highlands above the canyon rim in the general region of the Grand Canyon, we would find formations younger than the Kaibab. From their fragmentary successions we could build a composite that would include red, brown, yellow, and gray sandstones, conglomerates, and shales that contain the famous petrified forest of tree trunks and, in some places, dinosaur remains.

The rocks of the Grand Canyon have many stories to tell: of the advance and retreat of the seas over the continent at this place, of the appearance and disappearance of different kinds of organisms, and of the different kinds of marine and terrestrial environments in which this remarkable variety of sediments was deposited. But one of the most important is the tale of time—the time that is represented by the rocks of the Canyon and the time that is recorded by the unconformities between so many of the formations. From the radioactive time scale, which is based upon the decay of radioactive elements in minerals, we know that the oldest formation, the Vishnu, is about 1400 to 1500 million years old, and the top of the Kaibab about 225 million years old. An enormous amount of time is represented by these rocks and unconformities.

Rocks as records of earth movements

Angular unconformities not only date erosion intervals, but they also record ancient earth movements. Beds below such unconformities were folded tilted, faulted, and uplifted before erosion produced the more-or-less even unconformable surfaces that we observe today. Erosion was, in turn, followed by additional earth movement, for only crustal subsidence could account for a change from erosion to further sedimentaion. Unconformities, therefore, are records of periods of mountain-building, even though the roots of the mountains are all we can see today. **Disconformities**—time gaps between two units whose bedding planes are parallel—are less dramatic, but they too imply the same general sequence of uplift, erosion, and subsidence.

There are other ways of telling a time sequence too. Though igneous rocks are not layered, as sediments are, they too have characteristics that place them in time. Igneous intrusions injected as a mobile magma may show sharp contacts with surrounding, or **country**, **rocks**. These contacts cut across and interrupt original structures in the country rock. Such cross-cutting intrusions are thus **discordant**. They are typified by thin sheets, called **dikes**, that may lie at any angle to bedding. Intrusions may show **concordant** contacts, as in the case of **sills**, which follow the bedding of the sediments into which they were intruded. The concordant and discordant field relations between igneous rock and adjacent sedimentary, metamorphic, and other igneous rocks can be used to date these formations in just the same way that Steno's laws of original horizontality and superposition can be used to figure out relative ages of sediments. Similarly, folds and faults can be fitted into time sequences as well.

Figure 2-21 is the record of a series of events that can be interpreted in only one way. Of course, once such a time-event jigsaw puzzle has been put together, as in this diagram, it looks easy. But sorting out the pieces and seeing how they fit can sorely tax the imagination of a field geologist. To sum up, the ordering of geologic events with respect to a relative time scale is based on interpretations of sedimentary successions, igneous field relations, such as cross-cutting, and tectonic deformation, such as folding and faulting and angular unconformity. By piecing together the information gleaned from the study of such field relations, geologists during the nineteenth century worked out the entire stratigraphic time scale.

Hutton and uniformitarianism

Although the reasoning used here in interpreting the Grand Canyon sequence may seem obviously correct, it was not until the close of the eighteenth century that geologists were ready to believe that there had been any evolution of the Earth's surface. Up to then they had struggled to find an explanation for the field relationships of rock formations consistent with the assumption that the Earth was created just as we see it today, with all of its river valleys, mountains, and plains placed where they are by God. The new way of looking at the Earth included the recognition that constant change takes place as geological forces modify the surface and the interior.

A Scottish gentleman farmer, James Hutton, led the way with a book carrying the bold title **Theory of the Earth with Proof and Illustration**, first presented to the Royal Society of Edinburgh in 1785. Hutton's greatness lies in his recognition of the cyclical nature of geological changes, and of the way in which ordinary processes, operating over long time intervals, can effect great changes. He reasoned from observation that rocks slowly decay and disintegrate under the action of water and air. This process—**weathering**—produces debris in the form of gravel, sand, and silt and furthers erosion of the land. Water and air also act to transport the debris, most of which ends up near or below sea level. The deposits are compacted, cemented, and ultimately become sedimentary rocks. At a later time, according to Hutton, subterranean heat and thermal expansion may produce an intrusion of igneous rock. The **plutonic** episode (named for Pluto, Greek god of the underworld) would be accompanied by upheaval of the sediments and deformation into folds and faults accompanying mountain-building, or **orogeny**. Marine sediments emerge as land, bringing deposition to a halt, and then erosion of the newly emerged highlands initiates the cycle all over again.

Hutton observed and learned from the modern counterparts of each stage of this cycle: mountains erode, rivers carry debris to the sea, ocean waves pound rocks, sands and muds settle to the bottom and then are buried on the sea floor. Because the physical and chemical laws that govern geological behavior don't change with time, one can, by studying processes in the present, infer the behavior of those in the past. Hutton, followed by Charles Lyell (Principles of Geology, 1830), used and publicized this **principle of uniformitarianism**. Uniformitarianism, as we understand it today, does not hold that the rates of geological processes or their precise nature had to be the same. Volcanism may have been more frequent in the past than it is now. Nevertheless, ancient volcanoes surely released gases and deposited ash layers and lava flows, just as modern ones do when they erupt. One reason geologists were so intent on studying the eruption of Mt. St. Helens was to learn how to interpret the deposits of ancient volcanoes.

Most of the concepts used in modern field interpretation go back to discoveries made by many late eighteenth-century and nineteenth-century geologists, but it was Hutton who first recognized that igneous bodies must be younger than the rocks they intrude. Hutton also pointed out that debris fragments in a sedimentary or igneous formation must have been derived from a parent rock older than the formation itself, and he was the first person to grasp the idea that a cycle of upheaval, erosion, subsidence, and sedimentation would show as an unconformity in the stratigraphic record.

With these principles established, nineteenth-century geologists opened a new era. The history contained in rock formations could at last be deciphered, and those who read it could travel backward in time to view ancient landscapes. It became possible to reconstruct the interrelations between mountains, oceans, climates, animals, and plants long since gone. By this time geography and geology had historical counterparts, paleogeography and paleogeology. Geology entered a period of discovery and glamour.

Evolution and time scale

In 1859 Charles Darwin's **On the Origin of Species by Means of Natural Selection** was published, and the theory of organic evolution was launched. Along with it was launched one of the great controversies in the history of science that continues even today. The ideas of evolution were denounced by many as monstrous and antireligious, if not downright silly. But under the leadership of Lyell and Thomas Huxley, geologists and biologists moved in a few years to acceptance of the theory. The concept of evolution had enormous impact, for its theoretical framework gave support to the idea that the time-related changes in fossil species could be used to set up a stratigraphic time scale[4].

Armed with a refined science of paleontology, based on Darwinian evolutionary theory, and with opportunities for travel afforded by the period of world exploration that accompanied European economic expansion and imperialism, geologists mapped the surface of the earth and fitted together what we now call the **Phanerozoic** (known) time scale. The names of the time periods are taken either from the geographic locality where the formations were best displayed or first studied or from some characteristic of the formations. For example, the Jurassic is named from the Jura Mountains of France and Switzerland, and the Carboniferous is named from the coal-bearing sedimentary rocks of Europe and North America.

Each time period of the stratigraphic time scale is represented by its appropriate system of rocks, and we differentiate time units, the **periods**, from time-rock units (the rocks that represent time), the **systems**. Each major unit is divided: the period into **epochs**, and the system into **series**. Epochs and series have geographic names, except for the older names of many of the epochs, which are simply called Upper, Middle, or Lower. Thus the Upper Jurassic Series comprises the rocks of the Upper Jurassic Epoch

[4]Darwin's new theory was actually discovered by both Darwin and Alfred Russell Wallace, a young, unknown naturalist working in the East Indies (Indonesia). Darwin had in 1857 explained his theory in a now-famous letter to the great American botanist Asa Gray at Harvard; Darwin's theory had been germinating in preliminary drafts for almost 20 years. In 1858 Wallace sent Darwin a manuscript that he had written in the midst of a bout with intermittent fever in the Molucca Islands. Darwin's reaction was close to panic, for Wallace's manuscript outlined his theory in essential details. On the advice of two friends, Charles Lyell, then the most famous geologist in England, and Joseph Hooker, a leading English botanist, Darwin and Wallace both presented their views before the Linnaean Society of London on July 1, 1858. Independent simultaneous discovery of an idea whose time has come, and the rush to publish it, is not a new development in science!

of time.

New words and expressions

rock section 岩石剖面 the Grand Canyon (美国)大峡谷
bedrock 基岩 outurop(s) 露头 cliff(s) 悬 崖 ， 崖
gorge 峡，峡谷
terrain(=terrene)地域，地体，地形 stratifieation 层理
bedding 层理、层面 sediment 沉积物 sandbar 砂坝
superposition 叠加 varvad clay 纹泥粘土，季节泥粘土
fossil(s) 化石 limestone 石灰岩 shale 页岩
sandstone 砂岩 reptile(s) 爬行类 mammal(s) 哺乳类
vertebrate 脊椎动物 dinosanur 恐龙 fern 蕨类植物
remain(s) 残余，化石 shark 鲨鱼 coal 煤
Home Sapiens 人（种） faunal 动物群 mapping (地质)填图
formation (地层)组 lithology 岩性 survey 调查，测量
canal 运河、渠道 Palaeozoic 古生代 Cambrian 寒武纪
Ordovician 奥陶纪 Silurian 志留纪 Devonian 泥盆纪
Carboniferous 石炭纪 Permian 二叠纪 Mesozoic 中生代
Triassic 三叠纪 Jurassic 侏罗纪 Cretaceous 白垩纪
Cenozoic 新生代 Paleogene 古近纪 Neogene 新近纪
Quaternary 第四纪 Pleistocene 更新世 Holocene 全新世
handlens 放大镜 igneous 火成岩的 magma 岩浆
crystalline 晶质的，结晶的 intrusive 侵入的 extrusive 喷出的
volcanic eruption 火山喷发 foliation 页理
metamorphic 变质的 angnlar unconfomity 角度不整合
fold 褶皱 metamorphism 变质作用
fault 断层 conformity 整合
disconformity 假整合 Trilobite 三叶虫 arthopods 节肢动物
paleontology 古生物 transgression 海侵
regression 海退 dike(s) 岩墙 sill(s) 岩床
terrestrial 陆地的，地球的 cross-bedding 交错层
conglomerate(s) 砾岩 rustal subsidenceC 地壳沉降
country rock 围岩 wearthering 风化作用
cemented 胶结的 plutonic 深成的
orogeny 造山运动 principle of uniformitarianism 均变论
geography 地理学 Phanerozoic 显生宙
era 代—Period 纪—epoch 世（时代）
group 界—System 系—series 统（地层）

II. Rocks and Minerals

Rocks and the minerals that make them up are the tangible record of geologic processes. The minerals of the Earth are understood in terms of their molecular architecture—the way their atoms are arranged in crystal structures. The kinds of atoms and their chemical bonding determine not only the crystal structures but the chemical and physical properties of minerals, all of which are used for their identification. Rocks are divided into the three major groups, igneous, metamorphic, and sedimentary, on the basis of origin. They are further subdivided within each group according to mineral composition and texture, which provide the data that allow us to interpret details of their origin.

Picking up pretty stones and showing or wearing them must go far back into human prehistory. The earliest records of practical use of stones, though, are of arrowheads and spear points made of flint (a sedimentary rock) or obsidian (volcanic glass) , both of which are hard materials that break with sharp edges. From the practical use of individual stones as tools, weapons, and decorations, it was a big step to the wholesale mining or quarrying of rocks and minerals for building, for making clay for pottery, and then for the ores that contain metals. Today mining is done so expertly and intensively that geologists have come to concern themselves with the exhaustion of the world's valuable mineral resources; minerals of economic significance and their reserves are covered in Chapter 22.

The materials of the Earth

Because of the many uses of rocks and minerals, we have a practical curiosity about where they are found and how they were formed: we want to be able to find more. Yet there are other reasons, too, for rocks, as we have seen, are the only records of how the Earth evolved, and they are an important guide to how the Earth works today. For this reason, **mineralogy**, the study of minerals, and **petrology**, the study of rocks, are important sub-fields of geology. Finally, there is the intrinsic interest in the extraordinary range of the mineral kingdom, with its immense variety of color, form, and texture. Minerals and rocks, after all, give us the marble and alabaster of sculpture, the jade of Eastern carvings, and the pigments used by Rembrandt.

What information do we want from rocks?

If the nature of rocks is a clue to many of the things we want to know about the Earth, how do we go about interpreting it? We need a key, just as ancient historians needed the Rosetta stone to crack the "code" of Egyptian hieroglyphics before they could read that part of human history First of all we want to find out just what the minerals are made up of and how the rock is put together from its constituent minerals. From its composition we should be able to say something about where the parent material came from and what it was like. What was the magma like? Or, what were the source rocks of a sediment? Or, what were the preexisting rocks that were heated and compressed to make a metamorphic rock? From the composition and the texture of the rock we should also be able to tell something of the pressures and temperatures at which the rock was formed by comparing these properties with the artificial rocks and minerals made in the laboratory.

In this chapter we cover the nature of the rocks and minerals that make up the Earth’s crust and mantle. We first explore the relation of rock to mineral. Next we show how the external appearance and properties of minerals are related to the way in which their fundamental building blocks—the atoms and ions of the chemical elements—are connected with each other in the internal architecture of crystals. With that picture in mind we can describe the mineralogy and textures of the three great classes of rocks.

Rocks are made of minerals

A rock is many things. It is a collection of the particular chemical elements that make it up. Those elements are not found randomly mixed in a rock, but they are distributed among an assemblage of minerals. A mineral is a solid chemical compound that is characterized by a definite composition or a restricted range of chemical compositions and by a specific, regular architecture of the atoms that make it up. Like all chemical compounds. Minerals are homogeneous: a mineral cannot be separated mechanically into different substances. Minerals make up a rock just as bricks make up a brick wall, in a great variety of arrangements. In coarse-grained rocks the minerals are large enough to be seen with the naked eye. In some rocks the minerals can be seen to have **crystal faces,** smooth planes bounded by sharp edges; in others, such as a typical sandstone, the minerals are in the form of fragments without faces. In fine-grained rocks, the individual mineral grains are so small that they can be seen only with a powerful magnifying glass, the hand lens that the field geologist carries. Some are so small that a microscope is needed to make them out.

On the basis of certain characteristics, particularly physical and chemical properties, several thousand minerals can be distinguished, each defined by its unique set of properties. For thousands of years, people who have used minerals—whether miners looking for iron ore minerals or artists looking for minerals to grind into pigments—have used simple physical and chemical tests to distinguish one from another. Color is one obvious characteristic. Differences in hardness were found to make it easy to distinguish between minerals that look similar. How minerals break apart, some showing smooth cleavage planes and others rough irregular fractures, proved to be a reliable way to identify certain minerals. Simple chemical tests were found useful in the field, such as dropping acid on a mineral suspected of being calcite ($CaCO_3$) to see whether the mineral would fizz as it dissolved, releasing carbon dioxide bubbles.

Early in the study of minerals it was realized that all grains or crystals of a mineral, like quartz, have just about the same qualities regardless of the kind of rock in which they are found. Some minerals, particularly those that have a more complex mixture of atoms, vary slightly in their properties, depending on their precise composition. A mineral like garnet, for example, has a number of varieties. Each variety has its own range of composition, such as the proportions of iron and other elements, and hence, its own set of properties.

Rocks are not as uniquely defined by their properties as minerals are. Because of the immense number of ways in which the thousands of minerals can be combined, the geologist is faced with a bewildering array of rock types. The only way to make order out of this array is to classify like with like and to sort out by general type. The major division of rocks into igneous, sedimentary, and metamorphic is just such an aid. Within each major division there are many groups and types. Using characteristic properties, we can divide the rock kingdom into several hundred general types, each with its own more-or-less distinctive earmarks.

Despite all of these numbers, a remarkable amount can be done by knowing even a small number of the most common minerals and rocks. In most parts of the world a field geologist can make an accurate geologic map by knowing only a few dozen major minerals and even fewer common rock types. This simplification is possible because most of the thousands of known minerals are either rare or unusual. In addition, many minerals can be lumped into groups. Thus the geologist who can recognize garnet will do well, even though a mineral sophisticate who can distinguish the many varieties of garnet by their slightly different chemical compositions might do better. Naturally, the more we can distinguish, the more the information gleaned, and the greater the power of our theories of explanation. That is why petrologists have to know a great deal about mineralogy.

Just how do we go about identifying minerals and explaining their origins from their characteristics? In the field we still use external form and other obvious physical properties. In modern laboratories, however, advanced instruments are used to learn the basic composition and atomic architecture of minerals. These are the underlying determinants of the other properties. From analysis of crystal structure and external form, we draw the best conclusions of origin. The next sections take us from the study of crystal faces to the explanation of physical properties in terms of the atomic arrangement.

Crystals: faces and symmetry

The regularity of crystal faces is the most striking feature of the external form of minerals, and for many years minerals were studied and identified mainly by analyzing their symmetry. In addition to his earlier contributions—enunciating the laws of stratigraphy and recognizing fossils—Steno wrote in 1669 that quartz crystals, wherever found, always show the same angle between similar crystal faces. By the late eighteenth century, his *constancy of interfacial angles* became accepted as a generality applicable to all minerals. By 1801 the major work of the great crystallographer René Haüy was accomplished, all in the midst of the great upheaval of the French Revolution. Haüy summarized the laws of *crystal symmetry*, the regularities of Crystal faces. That symmetry, we now know, is a manifestation of the symmetry of the arrangement of the atoms that make up the crystal. As a consequence of the work of Haüy and others, the early part of the nineteenth century was a time of intense study of the relation between the external forms of minerals and their chemical composition, a development that paralleled the great geological exploration of the Earth and the growth of the geological time scale.

Naming igneous rocks

As was stated previously, igneous rocks are most often classified, or grouped, on the basis of their texture and mineral composition. The various igneous textures result from different cooling histories, whereas the mineral composition of an igneous rock is the consequence of the chemical makeup of the parent magma and the environment of crystallization. As we might expect from the results of Bowen's work, minerals that crystallize under similar conditions are most often found together composing the same igneous rock. Hence, the classification of igneous rocks closely corresponds to Bowen's reaction series.

The first minerals to crystallize—calcium feldspar, pyroxene, and olivine—are high in iron, magnesium, or calcium, and low in silicon. Since basalt is a common rock with this mineral makeup, the term *basaltic* is often used to denote any rock having a similar mineral composition. Moreover, because basaltic rocks contain a high percentage of ferromagnesian minerals, geologists may also refer to them as *mafic* rocks (from *ma*gnesium and *F*e, the symbol for iron). Due to their iron content, mafic rocks are typically darker and slightly heavier than other igneous rocks commonly found at the earth's surface.

Among the last minerals to crystallize are potassium feldspar and quartz, the primary components of the abundant rock granite. Igneous rocks in which these two minerals predominate are said to have a ***granitic*** composition. Geologists also refer to granitic rocks as being ***felsic***, a term derived from ***feld*****spar** and ***silica*** (quartz), Intermediate igneous rocks contain minerals found near the middle of Bowen's reaction series. Amphibole and the intermediate plagioclase feldspars are the main constituents of this compositional group. We will refer to rocks that have a mineral makeup between that of granite and basalt as being ***andesitic***, after the rock andesite.

Although the rocks in each of these basic categories consist mainly of minerals located in a specific region of Bowen's reaction series, other constituents are usually present in lesser amounts. For example, granitic rocks are composed primarily of quartz and potassium feldspar (K feldspar), but may also contain muscovite, biotite, amphibole, and sodium feldspar (Na feldspar).

Thus far this discussion has focused on only three mineral compositions, yet it is important to note that gradations among these types also exist. For example, an abundant intrusive igneous rock called *granodiorite* has a mineral composition between that of granitic rocks and those with an andesitic composition. Another important igneous rock, ***peridotite***, contains mostly olivine and thus falls near the very beginning of Bowen's reaction series.

Since peridotite is composed almost entirely of ferromagnesian minerals, its chemical composition is often referred to as ***ultramafic***. Although ultramafic rocks are rarely observed on the earth's surface, peridotite is believed to be a major constituent of the upper mantle.

An important aspect of the mineral composition of igneous rocks is silica (SiO_2) content. Recall that most of the minerals in igneous rocks contain some silica. Typically, the silica content of crustal rocks ranges from a low of 50 percent in basaltic rocks to a high of about 70 percent in granitic rocks. The percentage of silica in igneous rocks actually varies in a systematic manner which parallels the abundance of the other elements. For example, rocks low in silica contain large amounts of calcium, iron, and magnesium. Consequently, the chemical makeup of an igneous rock can be inferred directly from its silica content. Further, the amount of silica present in magma strongly influences its behavior. Granitic magma, which has a high silica content, is viscous and exists as fluid at temperatures as low as a 800℃.

On the other hand, basaltic magmas are low in silica and generally fluid. Basaltic magmas are also extruded at much higher temperatures, often 1200℃ or higher.

Recall that igneous rocks are classified on the basis of their mineral composition and texture. Moreover, two rocks may have the same mineral constituents but have different textures and hence different names. For example, the coarse-grained intrusive rock granite has a fine-grained volcanic equivalent called rhyolite. Although These rocks are mineralogically the same, they have different textures and do not look at all alike.

Granitic rocks

Granite is perhaps the best known of all igneous rocks. This is partly because of its natural beauty, which is enhanced when it is polished, and partly because of its abundance. Slabs of polished granite are commonly used for tombstones and monuments and as building stones.

Granite is a phaneritic rock composed of up to 25 percent quartz and over 50 percent potassium feldspar and sodium-rich feldspar. The quartz crystals, which are roughly spherical in shape, are most often clear to light gray in color. In contrast to quartz, the feldspar crystals in granite are not as glassy, but rectangular in shape and generally salmon pink to white in color. Other common constituents of granite are muscovite and the dark silicates, particularly biotite and amphibole. Although the dark components of granite make up less than 20 percent of most samples, dark minerals appear to be more prominent than their percentage would indicate. In some granites, K feldspar is dominant and dark pink in color, so that the rock appears almost reddish. This variety is popular as a building stone However, most often the feldspar grains are white, so that when viewed at a distance granite appears light gray in color. Granite may also have a porphyritic texture, in which feldspar crystals a centimeter or more in length are scattered among a coarse-grained groundmass of quartz and amphibole.

Granite is often produced by the processes which generate mountains. Because granite is a by-product of mountain building and is very resistant to weathering and erosion, it frequently forms the core of eroded mountains. For example, Pikes Peak in the Rockies, Mount Rushmore in the Black Hills, the White Mountains of New Hampshire, Stone Mountain in Georgia, and Yosemite National Park in the Sierra Nevada are all areas where large quantities of granite are exposed at the surface. As we can see from these examples, granite is a very abundant rock. However, it has become common practice among geologists to apply the term *granite* to any coarse-grained intrusive rock composed predominantly of light silicate minerals. We will follow this practice for the sake of simplicity. The student should keep in mind that this use of the term *granite* covers rock having a range of mineral compositions.

Rbyolite is the volcanic equivalent of granite. Like granite, rhyolite is composed primarily of the light-colored silicates. This fact accounts for its color，which is usually buff to pink or occasionally very light gray. Rhyolite is usually aphanitic and frequently contains glassy tragments and voids indicating rapid cooling in a surface environment. In those instances when rhyolite contains phenocrysts, they are usually small and composed of either quartz or potassium feldspar. In contrast to granite, rhyolite is rather uncommon. Yellowstone Park is one well–known exception. Here rhyolitic lava flows and ash deposits of similar composition are widespread.

Obsidian is a dark-colored, glassy rock which usually forms when silica-rich lava is quenched quickly. In contrast to the orderly arrangement of ions that is characteristic of minerals, the ions in glass are unordered. Consequently, glassy rocks like obsidian are not complosed of minerals in the same sense as most other rocks.

Although usually black or reddish-brown in color, obsidian has a high silica content. Thus, its composition is more akin to the light igneous rocks such as granite than to the dark rocks of basaltic composition. By itself, silica is clear like window glass; the dark color results from the presence of metallic ions. If you examine a thin edge of a piece of obsidian, it will be nearly transparent. Because of its excellent conchoidal fracture, obsidian was a prized material from which the American Indians made arrowheads and cutting tools.

Pumice is a volcanic rock which, like obsidian, has a glassy texture,. Usually found with obsidian, pumice forms when large amounts of gas escape through lava to generate a gray, frothy mass. In some samples, the voids are quite noticeable, while in others, the pumice resembles fine shards of intertwined glass. Because of the large percentage of voids, many samples of pumice will float when placed in water, Oftentimes flow lines are visible in pumice, indicating some movement before solidification was complete. Moreover, pumice and obsidian often form in the same rock mass, where they exist in alternating layers.

Andesitic rocks

Andesite is a medium gray, fine-grained rock of volcanic origin. Its name comes from the Andes Mountains where numerous volcanoes are composed of this rock type. In addition of the volcanoes of the Andes, many of the volcanic structures encircling the Pacific Ocean are of andesitic composition. Andesite quite commonly exhibits a porphyritic texture. In these cases, the phenocrysts are often light, rectangular crystals of plagioclase feldspar or black, elongated hornblende crystals.

Diorite is a coarse-grained intrusive rock that looks somewhat similar to gray granite. However, it can be distinguished from granite by the absence of visible quartz crystals. The mineral makeup of diorite is primarily sodium-rich plagioclase and amphibole, with lesser amounts of biotite. Because the white feldspar grains and dark amphibole crystals are roughly equal in abundance, diorite has a “salt and pepper” appearance.

Basaltic rocks

Basalt is a very dark green to black, fine-grained volcanic rock composed primarily of pyroxene and calcium-rich feldspar, with lesser amounts of olivine and amphibole present. When porphyritic, basalt commonly contains small, light-colored calcium feldspar phenocrysts or glassy- appearing olivine phenocrysts embedded in a dark groundmass.

Basalt is the most common extrusive igneous rock. Many volcanic islands, such as the Hawaiian Islands and Iceland, are composed mainly of basalt. Further, the upper layers of the oceanic crust consist of basalt. In the United States, large portions of central Oregon and Washington were the sites of extensive basaltic outpourings. At some locations these once-fluid basaltic flows have accumulated to thickness approaching 2 kilometers.

Gabbro is the intrusive equivalent of basalt. Like basalt, it is very dark green to black in color and composed primarily of pyroxene and calcium-rich plagioclase. Although gabbro is not a common constituent of the continental crust, it undoubtedly makes up a significant percentage of the oceanic crust. Here large portions of the magma found in underground reservoirs that once fed basalt flows eventually solidified at depth to form gabbro.

Pyroclastic rocks

Pyroclastic rocks are those which form from fragments ejected during a volcanic eruption. One of the most common pyroclastic rocks, called *tuff*, is composed of tiny ash-sized fragments which were later cemented together. In situations where the ash particles remained hot enough to fuse, the rock is generally called *welded tuff*. Since welded tuffs consist of glass shards, their appearance may closely resemble that of pumice. Deposits of partially welded tuffs are easily quarried and used as a durable building material. Several villages in Cappadocia in central Turkey, which date as far back as the fourth century, have been carved into vertical cliffs composed of this material.

Pyroclastic rocks composed of particles larger than ash are called *volcanic breccia*. The particles in volcanic breccia can consist of streamlined fragments that solidified in air, blocks broken from the walls of the vent, crystals, and glass fragments. Unlike the other igneous rock names, the terms *tuff* and *volcanic breccia* do not denote mineral composition.

Occurrence of igneous rocks

Although volcanic eruptions can be among the most violent and spectacular events in nature and therefore worthy of detailed study, most magma is believed to be emplaced at depth. Thus, an understanding of intrusive igneous activity is as important to geologists as the study of volcanic events. The structures that result from the emplacement of igneous material at depth are called **plutons**. Since all plutons form out of our view beneath the earth's surface, they can be studied only after uplifting and erosion have exposed them. The challenge lies in reconstructing the events that generated these structures millions or even hundreds of millions of years ago.

For the sake of clarity, we have separated our discussions of volcanism and plutonic activity. Volcanism will be treated in the following chapter; here we will concentrate on plutonic activity. Keep in mind, however, that these diverse processes occur simultaneously and involve basically the same earth materials.

Nature of plutons

Plutons are known to occur in a great variety of sizes and shapes. Some of the most common types are illustrated in Figure 3.18. Notice that some of these structures have a tabular shape, while others are quite massive. Also, observe that some of these bodies cut across existing structures, such as the layering of sedimentary beds, whereas others form when magma is injected between sedimentary layers. Because of these differences, intrusive igneous bodies are generally classified according to their shape as either **tabular** or **massive** and by their orientation with respect to the country (host) rock. Plutons are said to be **discordant** if they cut across existing structures and **concordant** if they form parallel to the existing structures. Further, as we can see in Figure 3.18, plutons are closely associated with volcanic activity. The largest intrusive bodies are thought to be the remnants of magma chambers which fed ancient volcanoes.

Dikes Dikes are discordant masses that are produced when magma is injected into fractures. The force exerted by the emplaced magma can be great enough to further separate the walls of the fracture. Once crystallized, these tabular structures have thickness ranging from less than a centimeter to more than a kilometer. The largest have lengths of hundreds of kilometers. Most dikes, however, are a few meters thick and extend laterally for no more than a few kilometers. Dikes are often oriented vertically and represent pathways followed by molten rock which fed ancient lava flows. Some dikes end at depth; still others terminate at plutons.

Dikes may weather more slowly than the surrounding rock. When exposed, these dikes have the appearance of a wall as shown in Figure 3.19. Dikes are often found radiating, like spokes on a wheel, from an eroded volcanic neck. In these situations the active ascent of magma is thought to have generated stress fractures in the volcanic cone.

Sills sills are tabular plutons formed when magma is injected along sedimentary bedding surfaces. Horizontal sills are the most common, although all orientations, even vertical, are known to exist. Because of their relatively uniform thickness and large extent, sills are believed to form from very fluid lava. As we may expect, sills most often are composed of basaltic magma, which is typically quite fluid.

The emplacement of a sill requires that the overlying sedimentary rock be lifted to a height equal to the thickness of the sill. Although this seems to be a formidable task, it may require less energy than forcing the magma up the remaining distance to the surface. Consequently, sills form only at rather shallow depths where the pressure exerted by the weight of overlying strata is relatively low. Although sills are intruded between existing layers, they need not be concordant along their entire extent. Large sills frequently cut across sedimentary layers and resume their concordant nature at a higher level.

One of the largest and best-known sills in the United States is the Palisades sill, which is exposed along the west shore of the Hudson River in southeastern New York and northeastern New Jersey. This sill is about 300 meters thick and, due to its resistant nature, has formed an imposing cliff that can be seen easily from the opposite side of the Hudson. Because of its great thickness and subsequent slow rate of crystallization, the Palisades Sill provides geologists with an excellent example of fractional crystallization. The sill formed from magma rich in the minerals olivine, pyroxene, and plagioclase. Olivine, the first and heaviest of these minerals to crystallize, sank toward the bottom and makes up about 25 percent of the lower portion of the sill. By contrast, near the top of the sill, olivine represents only about one percent of the rock mass. Conversely, the lightest mineral of this group, plagioclase, floated toward the top and comprises nearly two-thirds of the upper portion of the sill. Examples such as the Palisades are important to geologists because they confirm the results obtained in the laboratory, where the actual conditions found in nature can only be approximated.

In many respects, sills closely resemble buried lava flows. Both are tabular and often exhibit *columnar jointing*. Further, because sills form in near-surface environments and may be only a few meters thick, the emplaced magma is often chilled quickly enough to generate a fine-grained texture. When attempts are made to reconstruct the geologic history of a region, it becomes important to differentiate between sills and buried lava flows. Fortunately, under close examination these two structures can be readily distinguished. The upper portion of a buried lava flow usually contains voids produced by escaping gases and only the roks beneath a lava flow show evidence of metamorphic alteration. Sills, on the other hand, form when magma has been forcefully intruded between sedimentary layers. Inclusions of adjacent country rock found within the upper and lower zones of the structure and “baked” zones above and below are trademarks of a sill.

Laccoliths Laccoliths are similar to sills because they form when magma is intruded between sedimentary layers in a near-surface environment. However, unlike sills, the magma that generates laccoliths is believed to be quite viscous. This thick, nonfluid magma collects as a lens-shaped mass that arches the overlying strata upward. Consequently, a laccolith can be detected because of the dome it creates at the surface even before the overlying rock is stripped away by erosional forces.

Most large laccoliths are probably not much wider than a few kilometers. The Henry Mountains in southeastern Utah are composed of several large laccoliths believed to have been fed by a much larger magma body emplaced nearby. Some geologists also consider the well-known structure called Devil’s Tower, located in eastern Wyoming, to be the remnant of a laccolith.

Batholiths By far the largest intrusive igneous bodies are batholiths. The largest batholiths are linear structures several hundred kilometers long and nearly one hundred kilometers wide as shown in Figure 3.22. The Idaho batholith, for example, encompasses an area of more than 40,000 square kilometers. Indirect evidence gathered from gravitational studies indicates that batholiths are also very thick, possibly even extending through much of the crust. Based on the amount exposed by erosion, some batholiths are at least several kilometers thick. By definition, a plutonic mass must have a surface exposure of over 100 square kilometers (40 square miles) to be considered a batholith. Smaller plutons of this type are termed **stocks**. Many stocks appear to be portions of batholiths that are not yet fully exposed.

Batholiths are usually composed of rock types having chemical compositions near the granitic end of the spectrum, although diorite is also found. Small batholiths can be rather simple structures composed almost entirely of one rock type. However, Studies of large batholiths have shown that they resulted from several distinct events that occurred over a period of millions of years. The plutonic activity which created the Sierra Nevada batholith, for example, is thought to have occurred as five separate events over a 130-million-year period which ended about 80 million years ago during the Cretaceous Period.

Batholiths frequently compose the core of mountain systems. Here uplifting and erosion have removed the surrounding rock, thereby exposing the resistant igneous body. Some of the highest mountain peaks, such as Mount Whitney in the Sierra Nevada, are carved from such a granitic mass. Large expanses of granitic rock are also exposed in the stable interiors of the continents, such as the Canadian Shield of North America. These relatively flat outcrops are believed to be the remnants of ancient mountains that have long since been leveled by erosion. Thus, the rocks composing the batholiths of youthful mountain ranges were generated near the top of a magma chamber, whereas in shield areas, the roots of former mountains and the lower portions of batholiths are exposed. We will consider the role of igneous activity as it relates to mountain building further in Chaper 20.

Emplacement of batholiths

One ongoing and interesting debate in geology concerns the emplacement of granitic batholiths. One group of geologists supports the idea that batholiths formed from magma that migrated upward from great depths. This idea, however, presents a space problem. What happened to the rock originally in the location now occupied by these igneous masses? Further, the problem of explaining how magma is able to force its way through several kilometers of solid rock also plagued those supporting the magmatic origin of batholiths. The group opposing the magmatic origin hypothesis suggested that the granite in batholiths originated when hot ion-rich fluids and gases migrated through sedimentary rock and chemically altered the rock’s composition. This essentially metamorphic process of converting country rock into granite without passing through a molten stage is called **granitization**. Although granitization undoubtedly generates small quantities of granite, the strongest evidence points to a magmatic origin for the largest intrusive bodies.

This controversy was resolved, for many at least, when careful studies were made of structures called *salt domes.* These structures are of economic importance as they are found in close association with major oil-producing areas in the Gulf Coast states and the Persian Gulf. Salt domes are produced in regions where extensive salt deposits were subsequently buried by thousands of meters of sediment. The salt, which is less dense than the overlying sediments, migrates very slowly upward. This is possible because salt behaves like a mobile fluid when it is subjected to differential stress over a long period of time. Since salt beds are not perfectly uniform, the zone of upward movement is thought to originate at a high spot along the layer. As the salt moves slowly upward the stress exerted on the overlying sediments causes them to mobilize and be pushed aside. Occasionally the salt breaches the surface, where it begins to flow outward not unlike a very thick lava flow.

It is now generally accepted that batholiths are emplaced in a manner similar to the formation of salt domes. Because magma is less dense than the overlying rock, its buoyancy propels it upward. Also, like a salt dome, the mobile magma forcibly makes room for itself by pushing aside the country rocks. As the magma moves upward, some of the country rock which was shouldered aside will fill in the space left by the magma body as it passes. An analogous situation occurs when a can of oil-base paint is left in storage. The oil in the paint is less dense than the pigments used for coloration; thus, oil collects into drops that slowly migrate upward while the heavier pigments settle to the bottom. In the case of ascending granitic magma, gradual cooling results in a loss of mobility. Thus, much of the magma crystallizes at depth to form granite batholiths rather than extruding at the surface as a volcanic eruption.

The upper portions of batholiths often contain unmelted remnants of the country rock that are called **xenoliths**. These inclusions indicate that yet another process may operate during the emplacement of a batholith, at least in a near-surface environment where rocks are brittle. As the magma buoys upward, stress is believed to cause numerous cracks in the overlying rock. The force of the injected magma is strong enough to dislodge blocks of the surrounding rock and incorporate them into the magma body. However, this process of assimilating country rock is only minor compared to the earlier mentioned activity of mobilizing and displacing country rock.

New words and expressions

minenal 矿物	mineraeogy 矿物学	sedimentary 沉积的
petrology 岩石学	flint 燧石	obsidian 黑曜岩
marble 大理石，大理岩	cleavage 解理	alabaster 雪花石膏
jade 玉	pigment 色素，颜料	garnet 石榴子石
crystallography 结晶学	clistallographer 结晶学家	calcium feldspar 钙长石
pyroxene 辉石	olivine 橄榄石	basalt 玄武岩
mafic 铁镁质的	potassium feldspar 钠长石	granite 花岗岩
granitic 花岗岩质的	felsic 长英质的	amphibole 闪石
plagioclase feldspar 斜长石	andesite 安山岩	andesitic 安山岩质的
muscovite 白云母	biotite 黑云母	granodiorite 花岗闪长岩
diorite 闪长岩	peridotite 橄榄岩	rhyolite 流纹岩
phaneritic 显晶质的	porphyritic 斑状	pumice 浮石, 泡沫岩
aphanitic 隐晶质的	conchoidal 贝壳状	hornblende 角闪石
phenocryst 斑晶	pyroclastic 火成碎屑的	tuff 凝灰岩
welded tuff 熔凝灰岩	volcanic breccia 火山角砾岩	stream lined 流线性的
pluton 侵入体、深成岩体	dike(s) 岩墙	sill(s) 岩床
batholiths 岩基	salt dome 盐丘,盐穹	buoyancy 浮力
xenolith(s) 捕容体，捕虏岩	collumnarjointing 柱状节理	laccoliths 岩盖

III. Volcanism and Plate Tectonics

The origin of magma has been a controversial topic in geology almost from the very beginning of the science. How do magmas of different compositions arise? Why do volcanoes located in the deep-ocean basins primarily extrude basaltic lava, whereas those adjacent to oceanic trenches extrude mainly andesitic lava? Why are basaltic lavas common at the earth's surface, whereas most granitic magma is emplaced at depth? Why does an area of igneous activity commonly called the "Ring of Fire" surround the Pacific Ocean? New insights gained from the theory of plate tectonics are providing some answers to these questions. We will first examine the origin of magma and then look at the global distribution of volcanic activity as viewed from the model provided by plate tectonics.

Origin of Magma

Based on available scientific evidence, the earth's crust and mantle are composed primarily of solid rock. Further, although the outer core is in a fluid state, this material is very dense and remains deep within the earth. What then is the source of magma that produces the earth's volcanic activity?

We know that magma can be produced when rock is heated to its melting point. In a near-surface environment, rocks of granitic composition begin to melt at temperatures around 800℃， whereas basaltic rocks must reach temperatures above 1000℃ before melting will begin. One important difference exists between the melting of a substance that consists of a single compound, such as ice, and the melting of igneous rocks, which are mixtures of several different minerals. Whereas ice melts at a definite temperature, most igneous rocks melt over a temperature range of a few hundred degrees. As a rock is heated, the first liquid to form will contain a higher percentage of the low-melting-point minerals than does the original rock. Should melting continue, the composition of the melt will steadily approach the overall composition of the rock from which it is derived. Most often, however, melting is not complete. This process, known as **partial melting**, produces most, if not all, magma.

A significant result of partial melting is the production of a melt with a higher silica content than the parent rock. Recall that basaltic rocks have a relatively low silica content and that granitic rocks have a much higher silica content. Consequently, magmas generated by partial melting are nearer the granitic end of the compositional spectrum than the parent material from which they formed. As we shall see, this idea will help us to understand the global distribution of the various types of volcanic activity.

What is the heat source to melt rock? One source is the heat liberated during the decay of radioactive elements that are thought to be concentrated in the upper mantle and crust. Workers in underground mines have long recognized that temperatures increase with depth. Although the rate of increase varies from place to place, it is thought to average between 20℃ and 30℃ per kilometer in the upper crust. This gradual increase in temperature with depth is known as the **geothermal gradient**.

If temperature were the only factor to determine whether or not a rock melts, the earth would be a molten ball covered with only a thin, solid outer shell. However, pressure also increases with depth. Since rock expands when heated, extra heat is needed to melt buried rocks, in order to overcome the effect of confining pressure. In general, an increase in the confining pressure causes an increase in the rock's melting point. Another important factor affecting the melting point of rock is its water content. Up to a point, the more water present, the lower the melting point. The effect of water on lowering the melting point is magnified by increased pressure. Consequently, "wet" rock under pressure has a much lower melting temperature than "dry" rock of the same composition. Therefore, in addition to a rock's composition, its temperature, confining pressure, and water content determine whether the rock exists as a solid or liquid.

In nature, rocks melt for one of two reasons. First, rocks melt when they are heated to their melting points. This could occur, for example, when hot mantle rocks migrate toward the surface and come into contact with crustal rocks having lower melting points. The volcanic activity in the region that includes Yellowstone National Park is believed to have resulted from such activity. Here basaltic magma that formed in the mantle transported heat into the crust, where partial melting of crustal rocks generated silica-rich magmas. Second, without increasing temperature, a reduction in the confining pressure can lower the melting temperature sufficiently to trigger melting. This occurs whenever rock ascends, thereby moving into a region of lower pressure. Both processes are thought to play significant roles in magma formation.

Most basaltic magmas are believed to originate from the partial melting of the rock peridotite, the major constituent of the upper mantle. Laboratory studies confirm that partial melting of this dry, silica poor rock produces magma having a basaltic composition. Since mantle rocks exist in environments that are characterized by high temperatures and pressures， melting often results from a reduction in confining pressure. This can occur, for example, where mantle rock ascends as part of a slow-moving convection cell.

Due to the fact that basaltic magmas form many kilometers below the surface, we might expect that most of this material would cool and crystallize before reaching the suface. However, as dry basaltic magma moves upward, the confining pressure steadily diminishes and further reduces the melting point. Basaltic magmas appear to ascend rapidly enough so that as they enter cooler environments the heat loss is offset by a drop in the melting point. Consequently, large outpourings of basaltic magmas are common on the earth's surface.

Conversely, granitic magmas are thought to be generated by partial melting of water-rich rocks that were subjected to increased temperature. As a wet granitic melt rises, the confining pressure decreases, which in turn reduces the effect of water on lowering the melting temperature. Further, granitic melts are high'in silica and thus more viscous than basaltic melts. Thus, in contrast to basaltic magmas that produce vast outpourings of lava, most granitic magmas lose their mobility before reaching the surface and therefore tend to produce large intrusive features such as batholiths. On those occasions when silica rich magmas reach the surface, explosive pyroclastic flows, such as those that produced the Yellowstone Plateau, are the rule.

Even though most magma is thought to be generated by partial melting, once formed the composition of a magma body can change dramatically with time. For example, as a magma body migrates upward, it may incorporate some of the surrounding country rock. As the country rock is assimilated, the composition of the magma is altered. Further, recall from our discussion of Bowen's reaction series that during solidification, magma often undergoes fractional crystallization. This produces a magma quite unlike the parent material. These processes account, at least in part, for the fact that a single volcano can extrude lavas with a wide range of chemical compositions.

Distribution of Igneous Activity

For many years geologists have realized that the global distribution of igneous activity is not random, but rather exhibits a very definite pattern. In particular, volcanoes that extrude mainly andesitic to granitic lavas are confined largely to continental margins or volcanic island chains located adjacent to deep-ocean trenches. By contrast, most volcanoes located within the ocean basins, such as those in Hawaii and Iceland, extrude lavas that are primarily of basaltic composition. Moreover, basaltic rocks are common in both oceanic and continental settings, whereas granitic rocks are rarely observed in oceanic regions. This pattern puzzled geologists until the development of the plate tectonics theory clarified the picture.

Most of the more than 600 active volcanoes that have been identified are located in the vicinity of convergent plate margins. Further, extensive volcanic activity occurs out of view along spreading centers of the oceanic ridge system. In this section we will examine three zones of volcanic activity and relate them to global tectonic activity. These active areas are found along the oceanic ridges, adjacent to ocean trenches, and within the plates themselves.

Spreading Center Volcanism

As stated earlier, perhaps the greatest volume of volcanic rock is produced along the oceanic ridge system where seafloor spreading is active. As the rigid lithosphere pulls apart, the pressure on the underlying rocks is lessened. This reduced pressure, in turn, lowers the melting point of the mantle rocks. Partial melting of these rocks(primarily peridotite) generates large quantities of basaltic magma that move upward to fill the newly formed cracks.

Some of the molten basalt reaches the ocean floor, where it produces extensive lava flows or occasionally grows into a volcanic pile. Sometimes this activity produces a volcanic cone that rises above sea level as the island of Surtsey did in 1963. Numerous submerged volcanic cones also dot the flanks of the ridge system and the adjacent deep-ocean floor. Many of these formed along the ridge crests and were moved away as new oceanic crust which was created by the seemingly unending process of sea-floor spreading.

Subduction Zone Volcanism

Recall that ocean trenches are sites where slabs of oceanic crust are bent and more downward into the upper mantle. When the descending oceanic plate reaches a depth of about 100 kilometers, partial melting of the water-rich ocean crust and perhaps the overlying mantle rocks takes place. The partial melting of these materials is thought to produce basaltic as well as andesitic magmas. After a sufficient quantity has formed, the magma migrates upward because it is less dense than the surrounding rocks. When subduction volcanism occurs in the ocean, a chain of volcanoes called an **island arc** is produced. Examples of volcanic island ares include the Aleutian Islands, the Tongan Islands, and the islands of Japan.

When subduction occurs beneath continental crust, the magma generated is altered before it solidifies, no matter whether it crystallizes at great depths or as lava extruded on the earth's surface. In particular, fractional crystallization and the assimilation of crustal fragments into the ascending molten body usually lead to magma exhibiting an andestitic to granitic composition. The volcanoes of the Andes Mounains, from which andesite obtains its name, are examples of this mechanism at work.

Many subduction volcanoes border the Pacific Basin. Because of this pattern, the region has come to be called the *Ring of Fire*. Here volcanism is associated with subduction and partial melting of the Pacific sea floor. As oceanic plates sink, they carry sediments and oceanic crust containing abundant water downward. Because water reduces the melting point of rock, it aids the melting process. Further, the presence of water contributes to the high gas content and explosive nature of volcanoes that make up the Ring of Fire. The volcanoes of the Cascade Range in the northwestern United States, including Mount St. Helens, Mount Rainer, and Mount Shasta, are all of this type

Intraplate volcanism

The processes that actually trigger volcanic activity within a rigid plate are difficult to establish. Activity such as in the Yellowstone region and other nearby areas produced rhyolitic pumice and ash flows, while extensive basaltic flows cover vast portions of our Northwest. Yet these rocks of greatly varying compositions actually overlie one another in several locations.

Since basaltic extrusions occur on the continents as well as within the ocean basins, the partial melting of mantle rocks is the most probable source for this activity. Recall that earth tremors indicate that the island of Hawaii does in fact tap the upper mantle. One proposal suggests that the source of some intraplate basaltic magma comes from plumes of rising magma called **hot spots.** Hot spots are unusually warm regions that originate deep within the earth's mantle. Here elevated temperatures produce a rising plume of molten rock which frequently initiates volcanism at the earth's surface . Hot spots are believed to be located beneath Hawaii and may have formerly existed beneath the Hawaii and may have formerly existed beneath the Columbia Plateau.

Generally lavas and ash of granitic composition are extruded from vents located landward of the continental margins. This suggests that remelting of the continental crust may be one of the mechanisms responsible for the formation of these silica rich magmas. But what mechanism causes large quantities of continental material to be melted? One proposal suggests that a thick segment of continental crust occasionally becomes situated over a plume of rising magma; that is, a hot spot. Rather than producing vast outpourings of basaltic lava as occurs at oceanic sites such as Hawaii, the magma from the rising plume is emplaced at depth. Here the incorporation and melting of the surrounding country rock, coupled with fractional crystallization, result in the formation of a secondary, silica-rich magma which slowly migrates upward. Continued hot spot activity supplies heat to the rising mass, thereby aiding its ascent. The activity in the Yellowstone region may have resulted from just this type of activity.

Although the plate tectonics theory has answered many of the questions which have plagued volcanologists for decades, many new questions have arisen; for example, why does sea-floor spreading occur in some areas and not others? How do hot spots originate? These are just two of many unanswered questions that are the subject of continuing scientific research.

The lost continent of Atlantics

One of the most popular legends of all time concerns the disappearance of the so-called continent of Atlantis. According to the accounts provided by Plato, an island empire named Atlantis disappeared beneath the sea in a single day and night. This event, which reportedly took place between $1500_{\text{B.C.}}$ and $1400_{\text{B.C.}}$ and caused the collapse of the Minoan civilization, is thought by many to have been the result of a cataclysmic volcanic eruption.

Research efforts in the eastern Mediterranean have provided evidence apparently linking Atlantis to the volcanic island of Santorin. Although once a majestic volcano, Santorin now consists of five islands located roughly midway between the island of Crete and Greece. Evidence collected from core samples taken around the remnants of Santorin revealed that a violent eruption occurred about $1500_{\text{B.C.}}$This eruption generated great volumes of ash and pumice that reached a maximum depth of 60 meters. Many nearby Minoan cities were buried and their ruins preserved beneath the ash.[11] Even as far away as Crete, enough ash fell to kill crops and possibly livestock grazing on the ash-laden plants. Following the ejection of this large quantity of material, Santorin collapsed, producing a caldera 14 kilometers across. The eruption and collapse of Santorin undoubtedly generated large destructive sea waves that caused widespread destruction to the coastal villages of Crete as well as those on the nearby islands to the north.

The connection between the eruption of Santorin and the disappearance of Atlantis is further supported by the fact that the Minoan civilization also disappeared between $1500_{\text{B.C.}}$ and $1400_{\text{B.C.}}$ About $1400_{\text{B.C.}}$ some of the Minoan traditions began to appear in the culture of Greece.

Most scholars agree that the eruption of Santorin contributed to the collapse of the Minoan civilization. Was this eruption the main cause of the dispersal of this great civilization or only one of many contributing factors? Was Santorin the island continent of Altantis described by Plato? Whatever the answers to these questions, clearly volcanism can dramatically change the course of human events.

[1] One of these buried cities, Akrotiri, on the island of Thera, has been excavated.

New words and expressions

plate tectonics 板块构造
convergent plate 聚敛型板块，聚合板块
subduction zone 俯冲带
plume(s) 热柱
caldera 破火山口
deep-ocean trench(es) 深海槽
geothermal gradient 地热梯度
lithosphere 岩石圈
island arc 岛弧
cataclysmic 灾难的
assimilate(v.t.) 吸收，同化
hot spots 热点

IV. Sedimentation and Sedimentary rock

Burial and Accumulation of Sediment

A bar in a river may exist for only a few months, from the high-water stage that formed it to the next one, which destroys it. Of all the sediments that are made day by day, only a small fraction is preserved to record geologic time. Sedimentary environments differ in the probability of preservation, and the differences are related, in part, to the rapidity of sedimentation and burial. In the deep sea, deposits accumulate at very low rates, only a few millimeters every thousand years. Once deposited, many of these deep-water sediments are unlikely to be eroded and redeposited because bottom currents strong enough for the task are unusual. That is why deep-sea pelagic sediments offer the most complete historical records of organic evolution, of temperature changes during ice ages, and of other geologic patterns.

In contrast, the shallower waters of the continental shelf deposit sediments at much higher rates—many centimeters every thousand years, on the average. These sediments are buried faster, but waves and currents are much more likely to scour the bottom and rework it mechanically. Furthermore, the higher the sedimentary accumulation, the more likely the waves are to rework and redistribute it; so that there is a level of sedimentation above which accumulation stops. The importance of **subsidence** of the crust is that it allows sedimentation to continue and the accumulation to build up. This process is one of slight positive feedback, in the sense that tectonic subsidence is reinforced by the sedimentary load. The further sedimentation proceeds, the greater the weight on the crust, because each layer of sediment is much denser than the sea water it replaces. The effect of the weight is to push down the crust (see the discussion of isostasy in Chapter 18), allowing still more sedimentation to proceed. This process does not continue endlessly, nor can it account for all of the subsidence shown by most continental shelves. The reason for this is that, as subsidence continues and sediments are laid down, more and more of the crust at that point consists of sediments; because these are much less dense than average crust, the effect of additional sediment tends to become negligible. Subsidence initiated by tectonic mechanisms and enhanced by sediment weight is evidenced by great thickness of sediments.

Geosynclines

The first to theorize about the origin of sediments of great thicknesses was the State Geologist of New York in the middle of the nineteenth century, James Hall. As a result of extensive field mapping of sediments of the Appalachian Mountains, correlating them stratigraphically, and measuring their thicknesses, Hall concluded that there were many tens of thousands of feet of sediment that had accumulated in a long, relatively narrow trough bordering the continent. He envisioned it as having been something like a large downfold in the surface that steadily accumulated sediment as it subsided. Later, the name geosyncline was applied to the sediment-filled trough.

For the past century the term has undergone a long and complex series of redefinitions tied to changing notions of the origins of the supposed troughs, mechanisms of subsidence in relation to sedimentation, the evolution of igneous and metamorphic rocks in the geosynclines, and of the nature of the structural deformation of these thick sedimentary sections. In the 1930s geosynclines were subdivided into an inner belt closer to the continental platform, the **miogeosyncline**, and an outer belt closer to an ocean basin, the **eugeosyncline**. The miogeosyn cline was characterized by lesser thicknesses of limestones, shales, and alluvial sandstones, though other kinds of sediment could also be found. The greater thicknesses of the sediments characteristic of the eugeosyncline included turbidite sandstones and shales, pelagic limestones and shales mixed with volcanic rocks, submarine basalts, and ash falls and flows.

An explosion of information on continental margins has accompanied the rise of plate-tectonic theory and its amplification in the last decade. This knowledge has led to new definitions, solidly based on geophysical and geological theory and observation of both oceans and continents. We now see geosynclines—there is still much of value in keeping the name—as thick, linear piles of sediments deposited along continental margins at present or former plate boundaries. The sandstones, shales, and carbonates of the continental shelves of passive continental margins accumulate to thicknesses of several kilometers as the edge of a rifted continent moves away from a mid-ocean ridge, cools, and subsides. These sediments correspond roughly to the miogeosyncline. The volcanics, turbidites, and forearc sediments of active continental margins are mixed with scraped-off pelagic sediments at subduction zones. This is the thicker sequence of the eugeosyncline. The two are brought together by a change in plate motions as a formerly passive continental margin converges with an active continental margin. When the two continents collide, their margins crumple first. Continental-rise turbidites of the one continent are mashed and faulted against the turbidites, volcanics, and pelagic sediments of the other. A collisional mountain chain is created of the thick sedimentary accumulations of the geosynclines.

Structurally deformed, metamorphosed, and injected with igneous intrusions, geosynclines were the forerunners of the major mountain chains of the world: the Alps, the Himalayas, the Appalachians, and the Western Cordilleran belt of the Americas. Only in the youngest mountain belts are the youngest geosynclinal beds preserved. As the mountains wear away, older and older sedimentary rocks are stripped off by erosion, in many places uncovering igneous and metamorphic rocks below. As geologic time goes on, the less likely it is that sediments uplifted into mountain ranges will survive erosion. In general, less and less sediment is found preserved as one moves back through the history of the Earth.

Continental platforms

Wide areas of the North American continent between the Appalachians and the Rocky Mountains have been relatively stable at least since the Cambrian Period began. Sedimentary accumulations are much thinner than in the geosynclines along the continental margins to the east and west, though there are a number of sedimentary basins, such as the Illinois Basin, that have received as much as 3000 meters (10,000 feet) of sediment. Deep-water formations are conspicuously absent, and many of the sediments are shallow carbonate-platform sediments and alluvial (or near-shore marine) sandstones and shales. Individual formations more or less continuous with those in geosynclines are much thinner on the platform. Subsidence has been much less on the platform, and no great mountain-making episodes have affected it. Conversely, erosion has not eaten so deeply into the surface sedimentary rocks, so that Precambrian igneous and metamorphic rocks, the basement, are exposed in only a few places, such as in the Ozark Mountains or the Black Hills. The key feature of platform sedimentation is not so much in sediment types and the environments in which they formed, for there is almost complete overlap with geosynclines in that regard, but in the thinness of the accumulation, a response to only mild and intermittent subsidence. The origin of this mild subsidence in areas far from plate boundaries remains in doubt.

Subsidence, which influences environment and thickness of accumulation, is only one aspect of the tectonic control of sedimentation. Just as important is the indirect effect that tectonics has on the mineral composition of detrital sediments.

Tectonics and sediment composition

Contrasts in the mineral composition of clastic sediments can be startling. For example, compare two alluvial sandstones, and arkose and a quartz arenite. The arkose may contain as much as 30 to 40 percent feldspar, the remainder being quartz and other mineral and rock fragments. The grains are coarse and angular, showing little sign of abrasion. The quartz sandstone may be more than 95 percent quartz with no feldspar. The grains are well sorted and well rounded. The contrast is not in the environment, for both are alluvial, but in the source of the erosional debris. The source of the arkose is likely a granitic igneous rock; the rocks eroded to make the quartz sand might well have been another sandstone.

As shown in Chapter 5, feldspar decomposes chemically to clay minerals, given enough time for reaction with rain and soil water. In comparison, quartz is essentially unaffected by most chemical weathering. The survival of feldspar (and thus the quartz-feldspar ratio) in a weathering terrain depends upon the rate of chemical decomposition in relation to the rate of mechanical erosion. That ratio is related to the topography of the terrain- the more elevated and rugged the mountains, the more rigorous the mechanical erosion. In relating the detrital quartz-feldspar ratio to mountainous topography, we have tied it to the product of tectonic activity. Tectonically quiet low-lying areas not subject to severe mechanical erosion produce detritus low in feldspar; much of the feldspar decomposes to clay before it can be transported.

This tectonic background to sediment composition is a major basis for the reconstruction of ancient tectonic events, of high mountain chains that once shed abundant detritus and are now gone. For example, the Devonian sandstones of the Catskill Mountains of New York are a very thick accumulation of alluvial and deltaic sands that were transported from a great mountain range to the east, roughly parallel to the present low hills of the Taconic Mountains east of the Hudson River. The mineralogy of the sandstones, including the types of feldspar, mica, other silicates and rock fragments, indicates a mixture of granitic igneous and high-grade metamorphic source rocks that are typical products of the mountain-building that results from continent-continent plate collision. Continental drift reconstruction shows this mountain range to have been formed by the junction of eastern and western continents at the beginning of the assembly of Pangaea. A similar analysis has been applied to the Tertiary formations of the San Joaquin Valley of California. Their mineralogy indicates volcanic island arc source rocks. The stratigraphic and structural relationships, coupled to plate tectonic reconstructions of the sea floor off California, show this sequence to have been formed in the fore-arc region of a subduction zone formed by convergence of a former oceanic plate, now entirely gone, and the North American plate. In Canada geologists have reconstructed fragmentary patterns of vanished mountain ranges, some more than three billion years old, from Precambrian sediments. In these ways, geologists have learned to use sediments as the hieroglyphics of geologic history.

Sediment into rock: diagenesis

Once a sediment is deposited and buried by other sediments, it is not immune to change. One has only to compare freshly deposited muds and sands with ancient shales and sandstones to see the obvious differences in hardness, cohesion, and porosity. The many processes that produce the changes in a rock's composition and texture after deposition are lumped together in the term **diagenesis**. Generally, they operate to harden the soft sediment into rock—that is, to **lithify** it. Diagenesis may also alter the mineral composition by dissolving some of the original minerals and precipitating new ones. The nature of oil, gas, and coal is almost completely the result of diagenesis of original sedimentary organic matter.

The major physical diagenetic change is **compaction**, a decrease in porosity caused by mineral grains being squeezed closer to each other by the weight of overlying sediment. Sands are fairly well packed as they are deposited, so they compact little, even when buried deeply. Newly deposited muds, however, are highly porous. They have a high water content when deposited, often well over 60 percent, and their layers thin drastically by compaction. The porosity equilibrium mixture of diverse minerals that have been brought together as detritus in the sediment (and, perhaps, mixed with some chemical precipitate from the environment, such as calcite). Thus, sedimentation may mix minerals from two very different kinds of igneous rocks, minerals that would have been incompatible under the original conditions of formation of the igneous rocks, such as a sodium-rich plagioclase feldspar from a granite and a calcium-rich one from a basalt. Diagenetic chemical changes tend to dissolve the calcium-rich feldspars and precipitate sodium-rich feldspars, thus moving the rock toward chemical equilibrium—namely, a more homogeneous plagioclase composition. A different example of lack of equilibrium is a grain of aragonite in a carbonate sediment which given enough time, tends to change to calcite, the form of calcium carbonate stable at low pressures and temperatures. This tendency to equilibrium results in many other chemical reactions between incompatible or unstable minerals, which result in the formation of new minerals and, thus, bring the rock to a new composition in equilibrium with its surroundings.

The second tendency is for a sediment to be buried more or less deeply in the crust. As a sediment is buried, it is subjected to increasingly high temperatures--on the average 1°C for each 30 meters (100 feet) of depth on the continents (see Chapter 13) - and high pressures - on the average about 1 atmosphere for each 4.4 meters of depth (1 pound per square inch for each foot of depth). As minerals and the surrounding groundwater in pore spaces are heated and put under greater pressure, they tend to react chemically to form new minerals. This process, when carried far enough, becomes metamorphism, in which the entire character of the rock alters. The boundary between diagenesis and metamorphism is somewhat arbitrary, usually drawn at a temperature of about 300℃.

The list of specific diagenetic changes is great. Few of these are universal; which ones occur depends on the geologic situation. A few important examples include the change of unstable opaline silica shells of diatoms to quartz, the stable form; transformation of the unstable swelling clay, smectite, to a stable mica type of clay, illite; the precipitation of calcite and quartz as pore-filling cement in sandstones; and the alteration of volcanic glass to clay minerals and zeolite minerals. All of these changes must be understood before we can deduce the nature of the original sediment that was laid down, as we must do in order to interpret geologic history.

Diagenesis is of practical value, because recovery of oil, gas or groundwater from formations depends on the porosity of the rock, which may have been largely determined by diagenetic cementation. The formation of oil itself is a diagenetic process by which organic remains of many kinds of organisms are gradually altered to liquid petroleum or natural gas, or both. The transformation of peat through bituminous coal to anthracite is another diagenetic process. Understanding all of these processes better enables us to search more intelligently and successfully for new mineral and energy resources.

New words and expressions

sediment 沉积物
deposit 沉积（物），矿床
burial 埋葬，埋藏
trough 槽，海槽，地堑
miogeosyncline 冒地槽，副地槽
turbidite 浊积岩，浊流沉积
alluvial 冲积的
isostasy 地壳均衡
passive continental margin 被动大陆边缘
collid (v), collision (n.) 碰撞，冲击
arkose 长石砂岩
abrasion 磨蚀（作用）
mica 云母
hardness 硬度
porosity 孔隙度
lithify (v.) 石化，使…成石
aragonite 霰石（文石）
opal 蛋白石

sedimentation 沉积作用
geosyncline 地向斜,地槽,沉降带
subsidence 下沉，沉降
downfold 向斜槽
eugeosyncline 优地槽，主地槽
topography 地形，地形学
pelagic 远洋的，深海的
continental shelf 大陆架
forearc 前弧
platform 台地，地台，平台
arenite 砂碎屑岩，砂岩
weathing 风化作用
hieroglyphics 象形文字
cohesion 粘性，凝结力
diagenesis 成岩作用
compaction 压实，致密
cementation 胶结作用
montmorillonite 蒙脱石

smectite 蒙脱石
illite 伊利石
bituminous 沥青质，沥青的
deltaic 三角洲的

bituminous coal 烟煤
peat 泥炭
anthractite 无烟煤

V. Plate Tectonics: a Review and Summary

Plate tectonics is the conceptual framework of this book, and we have already introduced the basic ideas in earlier chapters. In this chapter we draw together and review the diverse lines of evidence that support the theory of plate tectonics, using primarily illustrations you have already seen in earlier chapters. We will begin a discussion of rock associations and orogeny within the framework of plate tectonics. The fragmentation of Pangaea since the Jurassic will be reviewed together with some speculation about continental drift and extinct plates in the pre-Jurassic. The chapter will close with some brief remarks on the driving mechanism of plate tectonics, but the reader shouldn't expect more than vague speculations, for the subject is just beginning to receive serious study.

The mosaic of plates

According to the theory of plate tectonics, the rigid plates whose outlines are shown in the illustration inside the back cover slide over a partially molten, plastic asthenosphere in the general directions shown. According to the relative motions of adjacent plates, we can define three kinds of plate boundaries: (1) zones of divergence or spreading, typically ocean ridges; (2) fracture zones, or transform faults; and (3) zones of convergence.

Zones of divergence are boundaries along which plates separate. In the process of plate separation, partially molten mantle material upwells along linear ocean ridges, and new lithosphere is created along the trailing edges of the diverging plates. Such boundaries are characterized by active basaltic volcanism, shallow-focus earthquakes caused by tensile (stretching) stresses, and high rates of heat flow. The outpouring of magma along ocean ridges and the building of the oceanic lithosphere are volumetrically the most significant form of volcanism. Figures 1-13, 1-17, 13-8, 13-11, 15-19, and 15-37 emphasize the different aspects of divergence zones.

Typically transform faults are boundaries along which plates slide past one another, with neither creation nor destruction of lithosphere. Sometimes marked by scarps, transform faults are characterized by shallow-focus earthquakes with horizontal slips. Occasionally there occur "leaky" transforms, in which some volcanism and slight plate separation accompanies the transform. Examples are in Figures 1-17,17-13, and 18-17.

Zones of convergence are boundaries along which the leading edge of one plate overrides another, the overridden plate being subducted, or thrust into the mantle, where lithosphere is resorbed. The thrusting mechanisms that operate along these collision boundaries tend to produce deep-sea trenches, shallow- and deep- focus earthquakes, adjacent mountain ranges of folded rocks, and both basaltic and andesitic volcanism. Convergence zones are illustrated in Figures 1-16, 13-8, 13-11, and 15-37.

Each plate is bounded by some combination of these three kinds of zones, as can be seen on the inside of the back cover. For example, the Nazca Plate in the Pacific is bounded on three sides by zones of divergence, along which new lithosphere forms, and on one side by the Peru-Chili trench, where lithosphere is consumed. Continental margins may or may not coincide with plate boundaries. If they do, the continents tend to remain “afloat” because continental plates are thicker and too buoyant to be readily subducted. Where two plates with continents at their leading edges converge, the crust thickens to form great mountain ranges like the Himalayas.

The global sum of plate creation and consumption is approximately zero The Earth would otherwise change size in order to accommodate the new sea floor, and this doesn’s seem to be happening. Instead, the plates form and disappear and change in size and shape as they evolve.

The structure and evolution of plates

Figure 19-1 depicts some of the structural details of a rigid lithospheric slab, a plate, from its region of generation at a ridge axis to its region of subduction, where it is resorbed. Both oceanic and continental crusts cap the plate; the continent, embedded in the moving plate, is carried along passively by it. Thus, in a real sense, continental drift is simply a consequence of plate movements. Underneath is the plastic, partially molten asthenosphere-source of the raw materials that build new lithosphere. Once heated and partially melted, subducted lithosphere becomes a source of magma, which rises to feed the overlying volcanic chain. A generalized heat-flow profile. Shows a large amount of heat emerging along the ridge axis, less from the older, cooled slab, and more from the volcanic chain of the subduction zone and the marginal basin behind it, where a small region of secondary spreading occurs.

Geophysicists have made theoretical studies and computer models of the evolution of a plate, from its creation out of hot rising matter at ocean ridges, through its spreading and cooling phase to its subduction, with reheating, melting and final resorption in the underlying mantle. The models help explain some important geological and geophysical observations: the major features of the ocean floors, the variation of heat flow from the sea floor, the occurrence of volcanism at plate margins, and the location and mechanism of earthquakes in the subducted slab.

Ocean depths increase with age, t, of the sea floor in a remarkably simple manner. For the first 80 million years the data fit a curve in which ocean depth increases as $\sqrt{t}.$ This is precisely the relationship predicted if a plate cools and contracts as it spreads. Beyond 80 million years ocean depths tend to flatten out, compared to the theoretical cooling curve, as would be expected if a small amount of heat is flowing into the plate from the underlying hot asthenosphere.

When a cold plate is subducted, it remains cooler than the surrounding hot mantle for about 12 million years, only gradually warming as it penetrates more deeply. Slow-moving plates heat up and are assimilated at shallower depths, perhaps 400 kilometers (250 miles), than fast-moving plates, which can penetrate to about 700 kilometers before heating to the point of assimilation. The process of subduction involves very large forces, and in a general way these forces must be responsible for the deep-focus earthquakes that occur only in downgoing plates. The sudden failures associated with earthquakes take place until the plates become so warm that stress is relieved by slow plastic deformation rather than by faulting. This seems to be the likely explanation for the fact that no earthquakes occur below 700 kilometers.

Rates of plate motion

The velocities of moving plates are measured by dating ocean-floor magnetic anomalies (using the time scale of magnetic stratigraphy) and dividing the age of each anomaly into the distance between it and the ridge axis. The procedure was outlined graphically in the preceding chapter.

The worldwide pattern of sea-floor spreading is being worked out by using a combination of magnetic, seismic, and bathymetric data. The charts used earlier (Fig. 18-21 and inside back cover) map the world's zones of spreading, subduction, and fracture; their geographic locations were obtained from the positions of ocean ridges, deep-sea trenches, earthquake epicenters, and other indications of activity. On the basis of spreading rates determined from magnetic data, isochrones (contours that connect points of the same age) were drawn to show the age of the sea floor in millions of years. The distance from a ridge axis to a 50-million-year isochron, for example, indicates the extent of new ocean floor created in that period. In Figure 18-21, note the closer spacing of the isochrones in the Atlantic than in the Pacific, where the spreading rate is higher. Because the fracture zones offset the isochrones, the age of the sea floor changes abruptly as one crosses a fault. A summary of the rates and directions of plate motions, measured in centimeters per year relative velocity, is given in Figure 19-4. The largest spreading velocity, 18.3 centimeters per year, occurs between the Pacific and Nazca plates.

The first results announced by the Deep Sea Drilling Project represented a great triumph for the magneticians who worked out spreading rates. The goal of this joint project of major oceanographic institutions and the National Science Foundation was to drill through the sediments of the sea floor at many places in the world's oceans. Studying the sedimentary cores makes it possible to work out the history of the ocean basin directly, in contrast to the indirect methods of magnetic anomalies. Since sedimentation begins as soon as an ocean forms, the age of the oldest sediments in the core, those closest to the basaltic bedrock, dates the ocean floor at that spot. The age is obtained from the fossils found in the cores. No sediments older than about 150 million years have been found, attesting to the "youth" of the sea floor. The sediments become older with increasing distance from mid-ocean ridges, confirming the prediction of the sea-floor-spreading hypothesis. Figure 19-5 is a plot of the ages determined from drill cores from the Atlantic and Pacific Oceans against ages predicted from magnetic data. It is remarkable how closely the experimental points approach the straight line with slope of one, which represents perfect agreement. This agreement clinches the concept of magnetic stratigraphy and the hypothesis of sea-floor spreading.

As an interesting aside, we have included a photograph of the drilling vessel Glomar Challenger. It is 400 feet long, and amidships it carries a drilling derrick 140 feet high. The only ship of its kind in the world, it can lower drill pipe several kilometers to the sea floor and drill thousands of meters into the sediments and underlying volcanic rock. For the ship to accomplish such a feat required a technological breakthrough. A means had to be devised to hold the ship stationary, regardless of current, wind, or waves, during drilling. Otherwise, the drill pipe would break off. The problem was solved by developing a positioning device that uses sound waves from acoustic beacons planted on the sea floor. Any change in the ship's position is sensed by a computer that monitors the time of arrival of the sound pulses. The same computer controls bow and stern side thrusters and the ship's main propulsion to keep the vessel on station. The Glomar Challenger was the answer to those who said when lunar exploration started, "Better to explore the ocean's bottom than the backside of the Moon." We ended up doing both.

Geometry of plate motion

If the individual plates behave as rigid bodies, as seems a reasonable first assumption, several interesting and useful geometric consequences follow. By "rigid" we simply mean that the distances among three points on the same plate-say, New York, Miami, and Bermuda-do not change, no matter how the plate moves. But the distance between New York and Lisbon, of course, increases because the two cities are on different plates that are being separated along a narrow zone of spreading on the mid-Atlantic ridge. Listed here are some geometric principles, mostly self-evident, that govern the sliding of plates on a planet.

1. Along transform-fault boundaries, no overlap, buckling, or separation occurs; the two plates merely slide past each other without changing the surface area. Look for a transform fault if you want to deduce the direction of plate motions because the orientation of the fault is the direction of relative sliding of two plates, as Figures 1-17 and 17-13 show. Surface area obviously changes at zones of convergence or divergence where plates are subducted or created. The plates can move perpendicularly or obliquely to the trend of convergent boundaries, which are therefore not as reliable indicators of directions of movement as transform faults or divergence zones.

2. Magnetic anomaly stripes and isochrones are roughly parallel and are symmetrical with respect to the ridge axis along which they were created. Look at Figure 18-17 to see why this must be so. Since each magnetic strip or isochron marks the edge of an earlier plate margin, isochrones that are of the same age but on opposite sides of an ocean ridge can be brought together to show the positions of the plates and the configuration of the continents as they were in that earlier time. By this means we can reconstruct, for example, the opening of the Atlantic Ocean, as shown in Figure 19-7[5].

3. The point at which three plates meet is called **a triple junction.** Figure 19-8 shows an example of a point at which a spreading zone, a subduction zone, and a transform fault meet. If the relative motion between two pairs of plates is known, we can solve for the third by using a simple equation (see Box 19-1).

The point where the Pacific, Cocos, and Nazca plates meet (see inside back cover) is an actual triple junction. Three spreading zones meet at this junction, as shown in the enlarged view in Figure 19-9. The unknown motion, found by vector addition, was that between the Nazca and Pacific plates, the motions between the Pacific-Cocos and

[5] The Great Pyramid of Egypt is aimed slightly east of true north. Did the ancient Egyptian astronomers make a mistake in orienting the pyramid 40 centuries ago? Probably not. Over this period of time Africa drifted enough to rotate the pyramid out of alignment with true north.

Cocos-Nazca plates having been worked out from transform faults and magnetic-anomaly stripes. The arrows show the resultant plate movements. Note also how the isochron bend to become parallel to the spreading centers, where they originated, and how they are offset by the transform faults. The spacing of the isochron reflects the spreading rates, which are largest for the Pacific-Nazca plates and least for the Cocos-Nazca plates.

Up to this point we have considered plates sliding on a plane. Although much can be learned about plate motions by making this simplification, plates actually move on the Earth's spherical surface. Box 19-2 explains how plate movements on a sphere can be described. With the application of these geometric principles to find spreading directions and magnetic anomalies to deduce spreading rates, the relative motions of the lithospheric plates are being worked out worldwide. Some results have already been pictured in Figures 18-21 and 19-4. However, geophysicists are searching for ways to measure the absolute motions of individual plates rather than their motions relative to each other. If the hot spots discussed in Chapter 15 turn out to be fixed in the mantle below plates, then the string of extinct volcanoes trailing from the hot spot would record the movement of individual plates as they glide over the mantle. This is currently a subject of active research.

Sea-floor spreading and continental drift: rethinking Earth history

One of us (F. P.) once helped write a paper dealing with the permanence of ocean basins. If he were allowed to expunge from the scientific record the one contribution he regrets the most, this would be it. The notion of the stability of global geographic features was not only a main tenet of the old geology but seems to be firmly rooted in the human psyche. We now know that on the geological time scale the sea floor is far from permanent. The present ocean basins are being created by spreading and recycled by subduction on a time scale of about 200 million years, which is about 4 percent of the age of the Earth. The likelihood of finding extensive older remnants of sea floor is slight. Continents. On the other hand, are mobile but permanent features. They are too buoyant to be subducted. They may be fragmented, moved, reassembled, deformed, and eroded at their surfaces, but their bulk does not seem to be much diminished. Old terrains with ages of around 3.5 to 3.7 billion years can still be found. Continents grow with time by the gradual accumulation of materials along their margins. New continental strips can therefore be added on in different places at different times, depending on the history of fragmentation, movement, and reassembly.

With the emergence of these revolutionary ideas, geologists are rethinking Earth history. Most of the evidence for plate tectonics comes from the sea floor, a relatively simple place compared to the enormously complicated continents. Just how plate tectonics explains continental geology is now receiving much attention. New developments reported in nearly every issue of the geological journals show that the subject has definitely been revitalized. Rock associations, volcanism, metamorphism, the evolution of mountain chains—all are being reexamined in the framework of plate tectonics. Some of the new interpretations that we describe in this chapter may not stand the test of time. In this connection, future editions of this book may show some changes, not so much in the big picture of plate tectonics as in the details of fitting regional geology into the overall framework. The student (as well as the authors of this book) should be cautioned against calling on plate tectonics for easy explanations of everything geological. It is not clear, for example, how or whether the origin of such structures as the Ozarks, the Black Hills, the Colorado Plateau, or such intracontinental, sediment-filled depressions as the Michigan Basin are related to plate movement.

Rock assemblages and plate tectonics

The only record we have of past geologic events is the incomplete one found in the rocks that have survived erosion or subduction. Since only sea floor younger than 200 million years (the last 4 percent of Earth history) has survived subduction, we must focus on the continents to find the evidence for most of Earth history. Some of the methods of reading the rock record have been described in earlier chapters. Here we explore the nature of the rock assemblages that characterize different plate-tectonic regimes as a first step in unraveling the history of past plate motions. Our aim is to reconstruct the process of continent fragmentation and ocean development, to locate the sites of vanished oceans, and to recognize the sutures that mark ancient plate collisions.

Of the three kinds of plate boundaries, we might expect distinct suites (assemblages) of rocks to be associated with plate divergence and convergence. At transform faults no distinct or characteristic rock assemblages are to be expected. Discontinuities across the fault are found, however, since rock formations formed and altered elsewhere have slipped past one another, and once-continuous formations or structural features are displaced.

Think of all that happens at a zone of divergence, where plate accretion and spreading occur, and you can predict the kinds of rock that would characterize the place and the process. Because there is extensive undersea volcanism, one would expect to find submarine basaltic lava, perhaps pillow lavas, the volcanic rock formed when hot lava is quenched by cold sea water (Chapter 15). Suboceanic crust and mantle are created here; dredge hauls and geophysical data show these layers to consist of mafic rocks, such as gabbro and peridotite, often showing evidence of alteration in a water environment (hydrous metamorphism). A carpet of deep-sea sediments would cover all of this. From Chapters 10 and 11 we remember that these deposits are recognized by thin layers of shale, limestone, and the siliceous rock chert, often with thin, discontinuous turbidites between them. Some or all of these layers may contain fossil remains of open-ocean marine organisms. A combination of deep-sea sediments, submarine basaltic lavas, and mafic igneous instrusions like that shown in idealized section in Figure 19-10, is called an **ophiolite suite**. The presence of narrow ophiolite zones in convergence features like the Alpine-Himalayan belt and the Ural and Appalachian belts may indicate that slices of oceanic crust and mantle originally produced at accreting plate margins were thrust onto land when an ancient ocean finally disappeared as two continents converged. It is generally believed that the Appalachians, for example, mark the site at which the ancestral Atlantic Ocean (called Iapetus for one of the Greek gods) closed when North America and Africa converged about 375 million years ago. The Atlantic reopened a few hundred kilometers east of this old suture, about 200 million years ago, in a spreading episode that is still underway.

Continental-shelf deposits are sedimentary rock assemblages that are laid down in an orderly sequence under tectonically quiet conditions in a geosyncline at a receding continental margin. Figures 19-11 and 10-29 show the orderly sequence of deposits in the geosyncline that is still forming off the Atlantic coast of the United States. The continental margin there was formed when the American plate separated from the European plate about 200 million years ago. Resting on the offshore shelf is a wedge-shaped deposit of sediments eroded from the continent and carried into shallow water. Because the trailing edge of the continent slowly subsides as the spreading lithosphere cools and contracts, the geosyncline continues to receive sediments for a long time. The load of the growing mass of sediment further depresses the crust isostatically, so that the geosyncline can receive still more material from land. For every three meters of sediments received, the crust sinks two meters. The result of these two effects is that the geosynclinal deposits can accumulate in an orderly fashion to thicknesses of 10 kilometers or more. At the same time, the supply of sediments is sufficient to maintain the shallow-water environment of the geosyncline, or miogeosyncline, as we called it in Chapter 11.

The deposits show all of the characteristics of shallow-water conditions (Chapter 11). At the bottom of the entire sequence are rift valleys containing basaltic lavas and nonmarine deposits formed during the early stages of continental fissuring. In the early stages of shelf deposition, sandy materials started to fill the depression. Much was dropped on the continental slope, only to be moved later to the continental rise by turbidity currents. In deep water, very thick deposits can be built up in this way. As the shelf miogeosyncline builds up, deposition may become dominated by shales and carbonate platform deposits—indicators of a decrease in the supply of detritus from the continent.

Think what might happen to these geosynclines if the orderly, sequentially layered, gently dipping sediments were to become the leading edge of a plate in collision. In the following sections we describe some of the many possibilities.

Just as the events that take place in a convergence zone are different from divergence-zone phenomena, so do the rock assemblages have different characteristics. The main features of ocean-ocean or ocean-continent collision are shown in transverse section in Figure 19-12. Thick marine sediments, mostly turbidites, eroded from the continent or the island arc, rapidly fill the long marginal depressions. In descending, the cold oceanic slab stuffs the region below the inner wall of the trench with these sediments and with deep-sea materials brought with the incoming plate. Regions of this sort are enormously complex and highly variable, as they included turbidities and ophiolitic shreds scraped off the downgoing slab by the edge of the overriding plate—all highly folded, intricately sliced and metamorphosed. They are difficult to map in detail but recognizable by their distinctive mix of materials and structural features. Such a chaotic mess has been called a **mélange**. The metamorphism is the kind characteristic of high pressure and low temperature because the material may be carried relatively rapidly to depths as great as 30 kilometers, where recrystallization occurs in the environment of the cold slab. Somehow, perhaps by buoyancy and mountain-building, the material rises back to the surface much later. Find a mélange and you can't be too far from the place of downturn of an ancient plate, long since consumed, but leaving this relic of its existence.

Refer again to Figure 19-12. Parallel to the mélange is a magmatic belt that makes up the arcuate system of volcanoes, intrusions, and metamorphic rocks formed on the edge of the overriding plate. Here the conditions are dominated by the rise of magma from the descending plate. At the interface, where the descending plate slides past the overriding one, perhaps friction is great enough to melt the upper part of the downturned slab, including the subducted wet sediments and ocean crust. The liquids rise buoyantly from depths of 100 to 200 kilometers to erupt and build the volcanic chains on the leading edges of plates. The characteristic igneous rocks produced are andesitic lavas and granitic intrusives. Island arcs, built up from the sea floor, may contain larger amounts of basalt; continental margins typically erupt rhyolitic ignimbrite and are intruded by granitic batholiths below. In contrast to that in a mélange, the metamorphism in the magmatic belts is typically the result of recrystallization under conditions of high temperatures and low pressures. This is because the hot fluids rise close to the surface, delivering much heat to a low-pressure environment.

Paired belts of mélange and magmatism are the signatures of subduction. The essential elements of these features of collision have been found in many places in the geologic record. One can see mélange in the Franciscan Formation of the California Coast Ranges and magmatism in the parallel belt of the Sierra Nevada to the east. This paired belt marks the Mesozoic boundary between the colliding Pacific and American plates. It even shows the polarity of the convergence by the location of mélange on the west and magmatism on the east; the Pacific plate was the subducted one. Other paired belts—for example, in Japan—can be found along the continental margins framing the Pacific basin. The central Alps, a European example, were produced by the convergence of a Mediterranean plate with the European continent.

Seismic reflection profiles are beginning to provide "x-ray" views of layers deep within the crust. Figure 19-14, a remarkable example of this new technique, shows the Australian plate being subducted under the Eurasian plate at the Jave trench.

Orogeny and plate tectonics

Orogeny means mountain-making, particularly by folding and thrusting of rock layers. In the framework of plate tectonics, orogeny occurs primarily at the boundaries of colliding plates, where marginal sedimentary deposits are crumpled and magmatism and volcanism are initiated.

Consider first some scenarios of plate convergence. In Figure 19-15a, a plate with a continent at the leading edge collides with another plate carrying a continent. In the early stage, during which the convergence is between continent and subducted oceanic lithosphere, a magmatic belt, folded mountains, and mélange deposits may be features of the overriding continental boundary. An example exists today along the Pacific coast of South America, where the American and Nazca plates are colliding. Look at the illustration inside the back cover to see the setting of the plates. The Andes, from which the name of the volcanic rock andesite is derived, lie in the magmatic belt; subduction is taking place under the Peru-Chile trench.

In a later stage, continent may meet continent, as shown in Figure 19-15b. Since continental crust is too light for much of it to be carried down, the plate motions could be slowed or halted. Another possibility, the one depicted in the figure, is that the plate motions continue, with subduction ceasing at the continent-continent suture but starting up anew elsewhere. Cold and dense as the descending slab is, chunks of it may break off, fall freely into the mantle, and be resorbed. As Figure 19-15c shows, the suture is marked by a mountain range made up of either folded or thrust rocks, or both, coincident with or adjacent to the magmatic belt, and by a much-thickened continental crust. A prime example of continent-continent collision is the Himalayas, which began forming some 25 million years ago when a plate carrying India ran into the Asiatic plate (the collision and uplift are still going on). This may be how the root underlying the Himalayas originated (see Chapter 18). The plate-tectonic cycle of the closing of an ocean basin, a continent-continent collision, and the formation of an intracontinental mountain belt has been called the **Wilson cycle**, after the Canadian geologist J. Tuzo Wilson, who first suggested the idea that an ancient ocean closed to form the Appalachian mountain belt and then reopened to form the present day Atlantic Ocean.

Displaced terrains

Geologists have come across blocks within continents whose rock sequences, fossils, and paleomagnetism are alien to their surroundings. The rock assemblages and the fossils indicate different environments and ages than the surrounding terrain, and the paleomagnetic poles imply that the block originated in a different latitude. These are now believed to be fragments of other continents or of ocean crust that were swept up and plastered onto a continent in the process of plate collisions and separations. Coastal New England and Newfoundland may be slices of Europe; parts of Alaska, British Columbia, and Nevada may have been scraped off Asia; and central Florida may be a fragment of Africa. Displaced terrains have also been found in Japan, Southeast Asia, China, and Siberia, but their original locations have yet to be worked out.

The grand reconstruction

At the close of the Paleozoic, some 250 million years ago, there was a single supercontinent Pangaea, stretching from pole to pole. The fragmentation of Pangaea as a result of plate tectonics and continental drift over Mesozoic and Cenozoic time to form the modern continents and oceans is documented in the well-preserved record of magnetic reversal stripes on the ocean floor. But what of the pre-Pangaean distribution of continents? What were their shapes and where were they located? There is growing evidence that Pangaea was formed by the collision of continental blocks——not the same continents we know today but continents that existed earlier in the Paleozoic. The ocean-floor record for this period has been destroyed by subduction, so we must rely on the older evidence preserved on continents to identify and chart the movements of these paleocontinents. Old mountain belts like the Appalachians and the Urals mark the collision boundaries of the paleocontinents. Rock assemblages there reveal ancient episodes of rifting and subduction. Rock types and fossils also indicate the distribution of shallow seas, glaciers, lowlands, mountains, and climatic conditions. Paleomagnetic data can be used to find the latitude and the north-south orientation of the paleocontinents. Latitudes can also be checked by paleoclimatic data. Although it is not possible to assign longitudinal position to the paleocontinents, the relative sequence of continents around the globe can be pieced together from the fossil record. One of the first efforts to depict the per-Pangaean configuration of continents using these methods is shown in Figure 19-16. The ability of modern science to recover the geography of this strange world of hundreds of millions of years ago is truly impressive. Geologists may be able to continue to sort out more details of this complex jigsaw puzzle, whose individual pieces change shape over geologic time.

Figure 19-17 reconstructs the most recent breakup of Pangaea as we now understand it. Figure 19-17a shows the world as it looked in Permian times, a little more than 200 million years ago. Pangaea was an irregularly shaped land mass surrounded by a universal ocean called Panthalassa, the ancestral Pacific. The Tethys Sea, between Africa and Eurasia, was the ancestor of part of the Mediterranean. The fit of North and South America with Europe and Africa is very good in detail when taken at the outer edge of the continental shelves, instead of at the present shorelines, which are some distance from the original rift. It is the fit for which we have the firmest evidence. The positions of Central America, India, Australia, and Antarctica are less certain.

The breakup of Pangaea was signaled by the opening of rifts from which basalt poured. Relics of this great event can be found today in the Triassic basalt flows all over New England. Radioactive dating of these flows provides the estimate of about 200 million years for the beginning of drift.

The geography of the world after 20 million years of drift—at the end of the Triassic some 180 million years ago—is sketched in Figure 19-17b. The Atlantic has opened, the Tethys has contracted, and the northen continents (Laurasia) have all but split away from the southern continents (Gondwana). New ocean floor has also separated Antarctica-Australia from Africa-South America. India is off on a trip to the north.

By the end of the Jurassic period, 135 million years ago, drift had been underway for 65 million years. The big event at this time is the splitting of South America from Africa, which signals the birth of the South Atlantic. The North Atlantic and Indian oceans are enlarged, but the Tethys Sea continues to close. India continues its northward journey.

The close of the Cretaceous period 65 million years ago sees a widened South Atlantic, the splitting of Madagascar from Africa, and the close of the Tethys to form an inland sea, the Mediterranean. After 135 million years of drift, the modern configuration of continents becomes discernible.

The modern world, produced over the past 65 million years, is shown in Figure 19-17e. India has collided with Asia, bringing its trip to an end. Australia has separated from Antarctica. Nearly half of the present-day ocean floor was created in this period. Figure 19-18 shows several schematic sections that summarize modern plate, ocean, continent, and island-arc relationships for the American, African, Eurasian, and Indian plates.

Most of the modern Pacific Ocean basin consists of the Pacific plate side of the East Pacific rise spreading zone, as can be seen inside the back cover and in Figure 19-18b. This implies that an area equal to most of the Pacific Ocean has disappeared by subduction under the Americas in the past 130 million years. As much as 7000 kilometers (4300 miles) of Pacific sea floor may have been thrust under North America!

Not one branch of geology, except perhaps crystallography, remains untouched by this grand reconstruction of the continents. Economic geologists are using the fit of the continents to find mineral and oil deposits by correlating the formations in which they occur on one continent with their predrift continuations on another continent. Paleontologists are rethinking some aspects of evolution in the light of continental drift. For example, during most of the age of reptiles, the continents we know today were grouped together in two supercontinents, Laurasia and Gondwanaland. There continents were fragmented during most of the age of mammals, with faunas developing on the daughter continents isolated from one another. Is this why mammals diversified into so many more orders than the reptiles did, and in a much shorter period of time? Structural geologists and petrologists are extending their sights from regional mapping to the world picture, for the concept of plate tectonics provides the means of interpreting such geological processes as sedimentation and orogeny in global terms. For example, the Caledonian mountain belt that runs along the northwest margin of Europe is the predrift continuation of the Appalachian belt, and the trend of the Andes may be followed into Antarctica and Australia, as Figure 19-19 shows.

Oceanographers are reconstructing currents as they might have existed in the ancestral oceans, to understand better the modern circulation and to account for the variations in deep-sea sediments. Paleoclimatologists are “forecasting” backward in time to describe temperature, winds, the extent of continental glaciers, and the level of the sea as they were in predrift times. What better testimony to the triumph of this once-outrageous hypothesis than its ability to revitalize and shed light on so many diverse topics!

The driving mechanism of plate tectonics

Up to this point everything we have discussed might be categorized as descriptive plate tectonics. The geometry and rates of plate motions, the consequences of plate separation and collision have been described. But what drives it all? We will not fully understand plate tectonics until we can answer this question. The International Geodynamics Project enlisted the efforts of thousands of scientists in seeking the underlying cause of plate motions.

It is generally accepted that most of the mantle is a hot solid, capable of flowing like a liquid at a speed of about a centimeter per year, about the rate your fingernails grow. The lithosphere is broken into rigid plates, somehow responsive in their motions to the flow in the underlying mantle.

As is generally the case when there is an abundance of data in search of a theory, many hypotheses have been advanced. Some would have plates pushed by the weight of the ridges at the zones of spreading or pulled by the heavy down going slab at subduction zones. Others hold that the plates are dragged along by currents in the underlying asthenosphere. Figure 19-20 shows some of these ideas. In line with the discussion in Chapter 13, we agree with those who view the process not in piecemeal but as a highly complex convective flow, involving rising, hot, partially molten materials and sinking, cool, solid materials, under a variety of conditions ranging from melting to solidification and remelting. A significant part of the mantle must be involved, for slasbs are known to penetrate to depths of some 700 kilometers before being completely resorbed. Figure 19-20c, shows one of the first computer models of the process-one that neglects some of the effects just mentioned, but that nevertheless accounts for many observations. A rising plume of hot material, heated from below, reaches the surface at a center of spreading. It moves away from the center, cools near the surface, and the cooled boundary becomes solid, strong lithosphere. Finally becoming heavier after it has cooled, the lithospheric slab sinks back into the mantle in a subduction zone, where it is reassimilated, to be heated and to rise again in the future. Another theory proposes that hot, narrow jet-like plumes rise from the bottom of the mantle, feed the growing plate, and drive it laterally away from spreading centers where the plumes mostly occur. These same plumes are evidenced at the surface by hot spots. Among the problems left to the next generation of Earth scientists is the incorporation of such important details as the shapes of plates, the history of their movements, and the formation and growth of continents into an explanation of the distribution of convective currents in time and space.

Summary

1. According to the theory of plate tectonics, the lithosphere is broken into about a dozen rigid, moving plates. Three types of plate boundaries are defined by the relative motion between plates: boundaries of divergence, boundaries of convergence, and transform faults.

2. In addition to earthquake belts, many large-scale geological features are associated with plate boundaries, such as narrow mountain belts and chains of volcanoes. Boundaries of convergence are recognized by deep-sea trenches, inclined earthquake belts, mountains and volcanoes, and paired belts of mélange and magmatism. The Andes Mountains and the trenches of the west coast of South America are modern examples. Divergent boundaries (for example, the mid-Atlantic ridge) typically show as seismic, volcanic, mid-ocean ridges. A characteristic deposit of this environment is the ophiolite suite. Transform faults, along which plates slide past one another, can be recognized by their topography, seismicity, and offsets in magnetic anomaly bands. Ancient convergences may show as old mountain belts, such as the Appalachians.

3. The age of the sea floor can be measured by means of magnetic-anomaly bands and the stratigraphy of magnetic reversals worked out on land. The procedure has been verified and extended by deep-sea drilling. Isochrons can now be drawn for most of the Atlantic and for large sections of the Pacific, enabling geologists to reconstruct the

history of opening and closing of these oceans. Based on this method and on geological and paleomagnetic data, the fragmentation of Pangaea over the last 200 million years can be sketched.

4. Although plate motions can now be described in some detail, the driving mechanism is still a puzzle. An attractive hypothesis proposes that the upper mantle is in a state of convection with hot material rising under divergence zones and cool material sinking in subduction zones, The plates, according to this model, would be the cooled, upper boundary region of the convection cell.

New words and expressions

mosaic of plates 板块拼接
asthenosphere 软流圈
transform fault 转换断层
convergence zone 聚合带（俯冲带）
divergence zone 背散带（扩散带）
lithosphere，岩石圈
scarp(s) 悬崖
leading edge 主动端
thrust 逆断层，(v.) 冲，插
buoyant 有浮力的
geophysicist 地球物理学家
magnetic anomaly 磁异常
seismic 地震的
magnetic_stratigraphy 磁性地层学
bathymetric 等深的
positioning device 定位装置
acoustic beacon 声纳
thruster（螺旋） 推进器
accretion 侧向加积
pillow lava 枕状熔岩
ophiolite suite 蛇绿岩套
mélange 混杂岩 (法语来源)
ignimbrite 熔结凝灰岩(熔灰岩）
Tethys 特提斯（古地中海）

VI. Interactions of Life with the Environment

Once metazoans evolved, the well-known fossil record began - a record of a host of evolutionary events, as new organisms arose by the interaction of the genetic material of each species with its environment, for the physical and chemical environment of the Earth's surface is the arena of natural selection that determines which species survive and which cease to exist. But the environment is itself acted upon by the organisms. The surface of the Earth, its atmosphere, and its oceans, have been profoundly affected by developments in organic evolution.

The shelled organisms did more than mark history with their fossil remains. In the oceans these organisms secreted enormous quantities of calcium carbonate, calcium phosphate, and silica to make their shells. As they died, new types of sediment were thus created, as were constructional features, such as reefs. In the mid-Paleozoic, the higher plants evolved and the land surface became vegetated with coniferous trees, ferns, and other early plant species. Lushly vegetated swamps became possible for the first time, producing another kind of biological sediment, coal. Vertebrate life had its origin in the early Paleozoic, beginning with the fishes, followed by the amphibians, and then the climactic evolution of the reptiles, the age of dinosaurs, in the Mesozoic. Mammals appeared in the Mesozoic too, but they reached dominance only after the dinosaurs disappeared. None of these vertebrates had any significant effect on Earth's physical or chemical environment; that was to come with the next development: the evolution of humankind. The genus *Homo* appeared on the scene only a few million years ago, and evolved to *Homo sapiens*, our *species*, only within the last million.

Paleontologists have long sought to tie major events of organic evolution to major geologic events. Hundreds of species of invertebrates vertebrates, and plants became extinct at various times in geologic history, the most massive extinctions coming at the ends of the Paleozoic and Mesozoic eras. What caused such wholesale disappearance of species? The assembly of the supercontinent of Pangaea in the late Paleozoic has been linked to the Permo-Triassic extinctions. As continents collided to form Pangaea, most of the expanse of shallow continental shelf surrounding each continent disappeared, leaving only one narrow perimeter around the supercontinent. During the Paleozoic the shelves had harbored the most productive biological communities. Climatic extremes, including glaciation of parts of what are now Africa, Australia, and South America, worked with the geographic constriction of shelves to create environmental stresses great enough to decimate many species. When Pangaea rifted apart, forming wide new expanses of hospitable shelves, the survivors founded the new stocks of the Mesozoic world.

A visitor from space is credited with the extinctions at the end of the Cretaceous in one of the newest geological mechanisms hypothesized. In 1980 a group of researchers from Berkeley, California led by Luis Alvarez, a physicist, and his son Walter, a geologist, announced the finding of extraordinarily high concentrations of the element iridium—thirty times normal—in marine clays deposited in several places exactly at the end of the Cretaceous. They ascribe the iridium and some other anomalous element concentrations to the impact of an asteroid of about ten kilometers in diameter. They hypothesize that it hit the Earth traveling at about 90,000 kilometers per hour, throwing tens of quadrillions of tons of pulverized rock up into the stratosphere. Because of this dust, they speculate, sunlight would have decreased to about 10 percent of full moonlight for several years, killing off plants on land and in the sea and leading to extinction of many species from dinosaurs to foraminifera. Though the chemical evidence for such an impact may be strong, paleontologists point out that different groups of organisms became extinct at different times in different places near the end of the Cretaceous, not instantaneously. Much research is being done to evaluate the implications of this provocative hypothesis and how well it explains the geological and paleontological record.

The major geologic event associated with human evolution and early history was, of course, the Pleistocene glaciation. Early humans learned first to take advantage of natural protected habitats, such as caves, for refuge from the snow and ice and cold. The next step was learning to build shelter. And that step ultimately led to our huge buildings, highways, dams, canals, and all of the other new "landforms" of the modern Earth. *Homo sapiens* is, of course, a natural species, and in that sense anything we do is "natural," but the landscape of our civilization goes so far beyond any other organism's work in modifying "nature" that our influence is clearly of a different order of magnitude. Our influence has had its negative aspects too, for inadvertently our modification of the environment to suit ourselves may have produced hazards as well as benefits.

Hazards to the environment

"Environment and ecology" sections of public libraries and bookstores are disturbing places to visit. Titles are made up of words like "crisis," "survival," "threat," and "danger." Perhaps more revealing is the use of such words as "frail" and "fragile" to describe the Earth. How do we reconcile the doomsday view with the geological evidence that the Earth is a dynamic global system that is remarkably stable in its steady-state equilibrium? To what extent have people suddenly wakened to the real possibility that we are so efficient at controlling our own environment that we may foolishly be tipping "the balance of nature" on a global scale, and to what extent have they simply become aware of what the Earth has been like all the time?

Some environmental hazards are local and regional; others are global. Landslides and earthquakes are natural and local phenomena. We describe, in Chapter 5, how poor engineering and construction practices can precipitate landslides and, in Chapter 17, how minor earthquakes can be provoked by pumping fluids at high pressures into subsurface formations. At the opposite extreme are the global hazards, such as the worldwide increase of carbon dioxide in the atmosphere that might influence the Earth's climate.

Many concerns for specific regions or types of environment are purely esthetic——the preservation of natural beauty, wilderness areas, and, in general, "unspoiled" landscape. To the extent that such concerns are based on uniqueness of the landscape, a knowledge of geology may inform our decision to preserve or modify any specific element of that landscape. We know that river gorges may be common in a variety of terrains, but there is little question that there is only one that has the magnificence of the Grand Canyon of the Colorado.

Environmental effects on health

The environmental hazards of greater importance are those to public health. Are the levels of lead in our drinking water or food much above the natural level, and, if so, are they potentially a medical danger? How widespread is mercury pollution, and how toxic are a few parts per billion of mercury in drinking water? In such matters the job of the scientist is dual. One is to measure the spread of a pollutant through the Earth-surface system of atmosphere, surface waters, biological communities, and sediments, and to determine how much human activities are altering that spread. The other is to answer a series of medical questions. What bodily damage may result from long exposure to low levels of toxic metals or other pollutants? At what levels do obvious disease symptoms appear? At the extreme, what are the lethal doses? And at the opposite extreme, we must consider the ill effects of deficiencies of some metals in our diet.

Most perplexing are the maps that show the incidence of certain diseases, from cancer to kidney stones, for each county or region of a country. These maps show that people living in certain areas are seemingly more prone to certain diseases than those living elsewhere. Does the incidence of a disease have to do with the kinds of people who settled in a region, or with their descendants, or does it have to do with the local geological environment that affects the water, food, dust, and anything else that may affect health?

Toxic metals

The toxic metals are some of the most serious potential offenders. Mercury, for example, which is known to be toxic, first hit the headlines some years ago when a Canadian graduate student discovered high levels of the metal in the tissues of lake fish. At about the same time we became aware that acute mercury poisoning was occurring in people living around several bays in Japan where shellfish, severely contaminated by mercury-laden industrial wastes, constituted a large portion of the diet. Mercury occurs in small amounts in many rocks- about 0.2 part per million in granite and less than half that amount in the average crustal rock. The mercury in rocks is steadily released in small amounts to natural waters by ordinary chemical weathering processes. Most natural waters contain only a few parts per billion and thus are harmless. Some part of the mercury in water is naturally converted to an organic form, methyl mercury, which is the form most harmful to organisms. Medical data indicate that chronic mercury poisoning may arise from high levels (many parts per million) of the metal dissolved in water, much of it as methyl mercury. At these levels, mercury affects the nervous system in hidden ways, with few well-defined symptoms shown for long periods of time after the exposure. For these reasons, the World Health Organization has recommended the maximum permissible mercury level for human food, including fish, at 0.5 part per million.

Enough work has been done on the circulation of mercury at the Earth’s surface to show how the biological world and the mercury in the physical environment interact. In the past fifty years, many thousands of tons of mercury have been mined for use in electrical equipment, chemical processing plants, and pesticides. We can look on this as an accelerated weathering process by which much more mercury than normal is being released from rocks. Though some of the mercury in some chemical processes is reused, a great deal of it escapes into natural waters or is vaporized into the atmosphere. From there, it is distributed to lakes, the ocean, and various sedimentary environments. A fraction of the mercury, converted to methyl mercury by bacteria, is ingested by organisms and accumulates in their tissues. As larger animals eat smaller ones, more of the metal accumulates in the larger ones, so that very large fish, such as tuna or swordfish, may contain relatively large concentrations, perhaps a few hundredths of a part per million. In waters that are polluted by industrial waste, the levels may be higher. Eventually, this material is absorbed by sedimentary particles, particularly the clays, and is buried out of reach of the biological system.

The serious question about mercury is, by just how much does the mercury in our waters exceed natural levels, and how widespread are such high levels? It is by no means clear that, in most places in the world, natural levels have been exceeded greatly. We know of a few places where industrial wastes have been uncontrolled and the pollution has reached dangerous levels; but we have only recently begun to pay much attention to these elements. We need to monitor unpolluted and polluted natural waters and their biological communities much more carefully. Like other toxic metals, mercury has always been present in the human diet at very low levels. Exactly what the range of those levels is - and by how much they can be exceeded without danger - remains to be determined. Minimum levels of some metals are required for good health, if, for fear of absorbing “poisonous” substances, we were to use only distilled water for drinking and food preparation, we might produce more harmful consequences than those caused by most impurities in water. We are, after all, adapted to a natural habitat in which small amounts of almost every element are found.

Other metals have the same general cycles as mercury. The most important, from a medical point of view, are lead, cadmium, arsenic, chromium, and nickel. Lead is particularly important, both because it is very toxic and because it has been dispersed throughout the atmosphere by automobile and industrial emissions. Two aspects of the problem are apparent: locally, in inner cities or near major highways, the lead levels from emissions may be very high compared to those in suburban or rural regions; and globally, there is evidence that the entire atmosphere is being charged with more and more lead. The United States has for some years been converting to lead-free gasoline for automobiles. If and when other countries follow and this source of atmospheric lead is cut down, the problem will disappear, for lead is rained out of the atmosphere very quickly.

Some geological materials are harmful to health as inhaled dust, especially by miners and industrial workers. *Silicosis* results from breathing quartz dust, *black lung* from coal dust, and *asbestosis* from asbestos minerals. A recent discovery of an extremely high rate of an unusual lung disease in a village in Turkey was linked to the abundance of a zeolite mineral, erionite, in the volcanic tuffs used locally to build houses. As we find out more about these kinds of diseases, we learn to be more prudent in the use of potentially hazardous materials.

All environments are self-cleaning by sedimentation. Sooner or later, the contaminants we worry about will settle out of the atmosphere, lakes, and oceans. That is small consolation to most of us because rates of sedimentation are so slow for most dissolved pollutants that, even if sources of excessive pollution are cut off, dangerously high levels may remain for a hundred years. In a particular environment, any pollutant tends toward a steady-state level that is determined by the balance between input rate from natural and human activity, dispersal through the environment, and sedimentation rates in that environment. Historical records show whether the system is out of balance, whether the levels are increasing because total input is greater than output. In the future monitoring may also show how levels decrease as artificial inputs are lowered and the natural output continues. Input-output relations are shown well by the carbon dioxide system.

Carbon dioxide and climate

The small amount of carbon dioxide in the atmosphere (a little over 320 parts per million) has a profound effect on our climate. The atmosphere is relatively transparent to the incoming visible rays of the Sun. Much of that radiation is absorbed at the Earth's surface and then reemitted as infrared, invisible long-wave rays that radiate back away from the surface. The atmosphere, however, is relatively opaque and impermeable to infrared rays because of the combined effect of clouds and carbon dioxide, which strongly absorbs the radiation instead of allowing it to escape into space. This absorbed radiation heats the atmosphere, which radiates heat back to the Earth's surface. This is called the "greenhouse effect," by analogy to the warming of greenhouses, whose glass is the barrier to heat loss. Any process that alters the amount of carbon dioxide in the atmosphere may conceivably affect Earth's climate.

Since the start of the industrial revolution, about the beginning of the nineteenth century, we have been pumping carbon dioxide into the atmosphere at an accelerating rate by our burning of coal, oil, and gas. The carbon dioxide level of the atmosphere has been increasing, as shown by systematic measurements in various places in the world. The amounts of carbon dioxide added to the atmosphere have also been calculated from the figures for fuel consumption. There is a pronounced discrepancy: the level in the atmosphere has not risen as much as would have been predicted by the additional supply. This suggests some loss or output from the atmosphere, something absorbing the extra carbon dioxide, moderating the effect of the increased input.

Much of the carbon dioxide that is "missing" from the atmosphere has been mixed with the oceans. Gas molecules of carbon dioxide in the air are in equilibrium with dissolved gas molecules in the water. As the concentration of gas in the air increases, there is a tendency toward reestablishing equilibrium: the water dissolves more gas, taking some of the excess from the air. In this way, the oceans are absorbing some of the carbon dioxide produced by the industrial revolution (about 50 percent) and keeping the atmosphere from departing much farther from its natural levels. Nevertheless, in spite of the ocean's moderating effect, carbon dioxide levels are expected to reach about

375 parts per million by the year 2000, a significant increase over 320 parts per million in 1970 and 295 parts per million in the middle of the nineteenth century. The possible increase in average global temperature as atmospheric carbon dioxide builds up is not great, because of many mediating effects such as the greater cloudiness produced by a rise in temperature, which tends to lower the radiation coming in from the Sun. But even small changes in temperature may have large climatic effects.

At the present rates of fossil fuel burning we may expect a doubling of the carbon dioxide level sometime in the next century. Calculations based on models of atmospheric and oceanic dynamics indicate such a carbon dioxide level may increase the global average surface temperature by 1.5 to 4℃ depending on the many uncertainties in the calculations. Such an increase in temperature might have drastic effects on the world's climate and weather patterns, shifting arid and temperate zones, possibly changing the frequency of droughts, and changing the distribution of water supplies. Agriculture might be profoundly affected. Even more important is the potential warming of glaciers, releasing meltwaters to the oceans. One mass of ice, the West Antarctic ice sheet, is particularly vulnerable to melting and breakup because its rock bed lies far below sea level. We have evidence of episodes of shrinkage and expansion of this sheet that affected sea level in the recent geologic past. If warmer climates and consequent warmer oceans in the next century should cause the West Antarctic ice sheet to disintegrate over a period of decades to centuries, sea levels would rise by three to six meters. Large parts of the world's coastal cities would be flooded, a disaster hard to imagine.

But calculations of this kind are subject to many uncertainties. The increase in carbon dioxide supplied to the atmosphere depends on the size of the world's population and its use of fossil fuel for energy. The rise in carbon dioxide concentration in the atmosphere depends on the degree of oceanic absorption and the balance between photosynthesis and respiration on land and in the sea. The climatic response to a rise in atmospheric carbon dioxide concentration is calculated using most imperfect models of the atmosphere-ocean dynamic system, and many of the quantities used in the models are poorly known. Complicating matters further, geologic processes unrelated to carbon dioxide in the atmosphere may affect climate over short times scales. One of the most uncertain factors is the role of dust. These tiny particles high in the stratosphere may reflect much sunlight away from the Earth and so decrease the available radiant energy that warms the surface. The amount of dust put into the atmosphere by human activity has become measurable, but at present it is still small compared to that of volcanic dust from major eruptions. The Earth's atmosphere, at least in historic times, has quickly recovered from a few relatively large eruptions as rainfall washed the dust out of the air. It is possible, however, that an exceptionally long string of big eruptions, such as those geologists know of in the past, could have an important long-term cooling effect. The injection of much additional dust by industrial and agricultural activities over the next decades could also have an effect.

The issue of climate change has deep implications for the social, political, and economic life of the next century. It is too important to be left to calculations that depend upon so many uncertainties. Levels of carbon dioxide and dust continue to be monitored by research laboratories and work on models of the atmosphere and ocean is being stepped up so that we may better predict what will happen in the future.

Using land with geological foresight

Life is not all hazards. Part of our concern for the environment stems from feelings that we may not be making the most sensible decisions about our use of land. A geologist can foresee the unhappy consequences of scalping a hillslope of soft, potentially waterlogged formations and building many small homes on the resulting unstable tract. The siting of highways, dams and reservoirs, strip and underground mines, large-sale tract housing, and many other enterprises is best not left to chance development, for the costs of unwise choices are too often borne by the community as a whole. Though zoning regulations on building are traditional in many urban areas, they rarely take geological realities into account.

Any construction has to be firm on its foundations, but that is only one aspect of the problem. Builders have to take into account what will happen to the foundation if water from many septic tanks waterlog the soil. Underground piping of gas, water, and sewage may be affected by slumping, landslides, or earthquakes. Engineering in areas of permafrost requires special precautions, as builders of the Alaska pipeline found. Sewage discharged from septic tanks may find its way into subsurface water supplies, contaminating them with dissolved chemical substances. Garbage disposal is an increasing problem as communities move against open dumps and uncontained burning of trash. One response, spreading the garbage and covering it with soil as sanitary landfill, may also result in groundwater contamination if the site is not well chosen. The formations below the landfill should be impermeable enough to prevent downward seepage into the water table. These problems are more severe in the disposal of toxic chemical wastes. Love Canal in Niagara Falls, New York, a shallow trench filled with such waste, came under scrutiny in the late 1970s as the possible cause of unusually high incidences of miscarriages and birth defects among those people living in the neighborhood of the dump.

The underground disposal of radioactive waste has become one of the most important issues in waste disposal. The main concern is the disposal of high-level waste from nuclear weapons manufacture and nuclear power reactors, but disposal of low-level radioactive waste from hospitals and laboratories also constitutes a growing problem. Most plans for the disposal of high-level waste call for incorporating the waste in a glass or other relatively insoluble canister encased in corrosion-resistant metal and buried a few hundred meters. The burial repository would be chosen for its ability to serve as a natural barrier to migration of any of the radioactive materials should they leak from the canister. Repositories in pure, dry salt beds or salt domes, granites, shales, or tuffs might serve to isolate the waste from possible leakage for at least several thousands of years and, it is hoped, for up to a million year.

The radioactive waste disposal issue is crucial to any large expansion of nuclear power plants. It is particularly critical for the planned development of fast breeder reactors, which produce the element plutonium. A major component of atomic bombs, plutonium is also one of the most toxic substances known. As little as a few thousandths of a gram (a few hundred-thousandths of an ounce) is enough to kill a person.

Other land-use planning must take larger regions of the land into account. Is it wise, for example, to plan for the urbanization of an area known to be the recharge of an underground aquifer that supplies drinking water to the population? Another question is one of surface-water storage. Builders are often tempted to drain marshes and swamps and build on the reclaimed land, but such places are important storage depots for water (see Chapter 6). Once the wetlands are drained and paved, the area is much more subject to flash flooding: because there is no opportunity for the water to soak into the ground, it runs off immediately.

There are many other choices to be made that involve evaluations of the suitability of land for agriculture, recreation, or transportation. Once the suitability of a given piece of land is determined, we must still face the priority problem: we must determine whether one use of the land is more important than another. Those priorities, such as designating land for recreation rather than building development, are socially and politically determined, but the choices among the alternatives have to be illuminated by geological knowledge. If we are to live with our decisions on development and utilization of land and resources for different purposes, they ought to be based on planning that is as intelligent as possible. The geological contribution is an understanding of how all contemporary surface processes are related, and how the crust of the Earth beneath has been patterned by past geological events.

New words and expressions

metazoa 后生动物	reef 礁
vegetation 植被	coniferous 针叶植物
fern 羊齿，蕨类植物	amphibians 两栖动物的，两栖类
invertebrates 无脊椎动物	supercontinent 超大陆
pangaea(Pangea) 泛古陆,联合大陆	
iridium 铱	stratosphere 平流层，同温度
foraminifera 有孔虫	glaciation 冰川作用
landslide 塌方，滑坡	earthguake 地震
methyl 甲基	pesticide 农药，杀虫剂
industrial waste 工业废物	asbestos 石棉
asbetosis 石棉沉着病	zeolite 沸石
erionite 毛沸石	waterlogged 浸满水的
reservoir 水库，储集层	slumping 滑坡，塌滑
permafrost 永久冻土	seepage 渗透，渗漏，油苗

VII. Geochemistry

Geochemistry of the Continental Mantle-Basaltic Rocks

Introduction

Bowen (1928) noted that ‘at 37-60km [one] encounter[s] material of a peridotitic character… and if this is true the only source of basaltic magma… lies in the peridotite zone, from which it must be produced by selective fusion. Moreover, he was one of the first to suggest that extraction of basalt from peridotite led to the peridotite being more barren’. Although there was a very acceptable idea towards the end of the twentieth century, these ideas were advanced at a time when some of his colleagues believed in basaltic layers deep in the Earth as a source for basaltic magmas (Daly 1926,1933).

Experimental petrology greatly advanced our understanding of the origin of basalts. Kuno(1960) linked the genesis of different magmas to depth and some of the earliest basalt-peridotite experiments (Yoder and Tilley 1962) studied the formation of melts to depths of 90km. O’Hara (1965, 1968) noted that melts in equilibrium with peridotite were different from so-called ‘primary magmas’ formed at depth and the need for fractionation of olivine became apparent. Green and Ringwood (1967) presented a comprehensive scheme for the production of tholeiitic and alkaline basaltic magmas in terms of polybaric partial melting of undepleted mantle. Soon the importance of CO_2 in mantle processes was realized and the stability of mantle carbonate was linked to the genesis of kimberlites and carbonatities (Eggler 1975, 1976, 1978; Wyllie 1987; Wallace and Green 1988; Green and Wallace 1988). In contrast, the compositions of basaltic magmas produced by melting of dry or volatile-free spinel lherzolite were perhaps best estimated from the experimental work of Takahashi and Kushiro (1983) and Fujii and Scarfe (1985). Throughout the 1970s and early 1980s most models of basalt genesis were based on the assumption that large degrees of melting were necessary to produce basaltic magmas from a peridotitic source and that small degree melts were unlikely to the easily mobilized. However, recent theoretical and experimental considerations indicate that small volume melts are mobile at mantle pressures and temperatures and, earlier models may need to be revised (McKenzie 1984, 1985; Watson and Brenan 1987; Brenan and Watson 1988; McKenzie and Bickle 1988; Hunter and McKenzie 1989; McKenzie 1989；Watson *et al.*, Chapter 6, this 1968) and Schilling (1971) pointed out the complementary geochemical relationship between oceanic tholeiites (depleted in large ion lithophile (LIL) and light rare earth (LRE) elements) and alkaline basalts (enriched in LIL and LRE elements). Gast (1968) proposed that the source of mid-ocean ridge basalt (MORB) had undergone a partial melting episode prior to extraction of MORB and be related the origin of alkaline (i.e. ocean island) and tholeiitic (i.e. mid-ocean ridge) basalts to variable degrees of melting of a heterogeneous mantle source (cf. McMenzie and Bickle 1988). Basalts within ocean basins that were not produced at ocean ridges were believed to be produced by hot-spot magmatism (Wilson 1965; Morgan 1971) and it was soon realized that ridge basalts were derived from shallow depleted asthenosphere and ocean island basalts were supplied by deeper mantle plumes (Schilling 1973; Hofmann and Hart 1975; Sun and Hanson 1975).

Continental alkaline volcanic rocks and mantle domains

Lewis (1888,1897) studied the matrix of diamond (blue ground) and concluded that it was a porphyritic peridotite of igneous eruptive character. Bonney (1899) prefaced an article on the parent rock of the diamond by stating that to much had been written on the occurrence of diamond by 1899 that brevity was the best introduction (Bonney 1899 and references therein). Despite that much of the next 50 years saw a rapid increase in the much of the database concerning the geology and mineralogy of kimberlites (Wagner 1914; Williams 1932). Wagner (1914) believed diamonds to have formed from kimberlitic magmas and noted that some pipes were barren of diamonds. He suggested that this might be due to a lack of carbon or carbon compounds in the deep-seated peridotite zone which he believed to the ultimate source of kimberlite. Wagner was also the first to note that diamondiferous kimberlites were erupted in stable geological terrains, a point that was re-investigated over 50 years later (Frantsesson 1968; frantsesson and Prokopchuk 1968; Dawson 1970). They noted that most diamondiferous kimberlites in southern Africa and Siberia were confined to the ancient cratons and that their abundance decreased into the surrounding mobile belts.

Probably because of the association of kimberlites with old crust enriched in radiogenic strontium the high $^{87}Sr/^{86}Sr$ ratios in kimberlites were believed to be due to: (1) incorporation of micro-xenoliths of crust (Powell 1966); (2) contamination with Sr derived from overlying limestones and shales (Brookins 1967); or (3) postemplacement alteration (Berg and Allsopp 1972). The underlying assumption at this time was that all mantlederived rocks had a source similar to oceanic basalts and as such should be isotopically identical. If their isotopic composition was different, then this was taken as evidence of crustal contamination and must constitute 'high level' interaction with the crust. This was accepted despite the fact that many 'altered' kimberlites had apparently lower $^{87}Sr/^{86}Sr$ ratios than fresh kimberlites had apparently lower $^{87}Sr/^{86}Sr$ rations than fresh kimberlites (Paul 1979; Demaiffe and Fieremans 1981). Whilst a source similar to oceanic basalts for basaltic kimberlites is consistent with much of the modern isotopic data (Kramers 1977; Basu and Tatsumoto 1980; Kramers *et al*. 1981)， not all kimberlites are genetically the same. In a landmark contribution, Smith (1983) demonstrated that basaltic or Group I kimberlites had a mantle source with the Sr, Nd, and Pb isotopic characteristics of ocean island basalts (OIB) and were therefore interpreted as asthenospheric melts. In contrast, micaceous or Group II kimberlites had Sr, Nd, and Pb isotopic characteristics similar to basalt-borne and kimberlite-borne xenoliths and were there fore different from oceanic basalts. These data were interpreted to mean that micaceous kimberlites had a significant source contribution from the subcontinental lithospheric mantle. The Group I and II terminology introduced by Smith (1983) is now more commonplace than the 'basaltic-micaceous kimberlite' terminology of Wagner (1914). The model of depleted asthenosphere (OIB)-enriched lithosphere interaction proposed by Smith (1983) was re-interpreted by McCulloch *et al*. (1983) who believed that the data was better explained by interaction of asthenospheric (MORB) melts and enriched lithospheric mantle. The retraction of micaceous diamondiferous kimberlites to the craton the dominant contribution from lithospheric mantle in their magmagenesis supports the contention that an enriched Archaean keel exists beneath the cratons.

Geochemical data for alkaline and tholeiitic basalts provide us with important constraints on the global distributions of mantle reservoirs. Many continental alkaline basalts (e.g. basanites, nephelinites, and alkali olivine basalts), like basaltic kimberlites, are chemically indistinguishable from ocean island basalts (Allegre *et al*. 1982), a fact that requires parts of the upper mantle to have been a common reservoir beneath both ocean basins and continents. Moreover, this reservoir must have been isolated for billions of years(Sun and Hanson 1975; Tatsumoto 1978; Sun 1980; Chase 1981). Whilst most investigators agreed that an enriched reservoir was required for alkaline ocean island and continental basalts and a depleted reservoir for midocean ridge basalts the exact geometry of these reservoirs remained a hot debated topic. Gast (1968) and Schilling (1971) proposed that the source for alkali basalts was deep and the source of MORB was shallow, Tatsumoto (1978) proposed the opposite with a shallow enriched reservoir and a deep depleted reservoir. Anderson (1979 *a*, *b*) queried a more fundamental assumption—that the upper mantle was necessarily peridotite. He believed that, as a result of early differentiation within the Earth, a layer of eclogite lay beneath a layer of peridotite. MORB was derived from this eclogite and the source was replenished by subduction.

To this day the exact geometry of MORB and OIB sources, the latter feeding continental and alkaline intraplate volcanoes, remains the source of some debate (Allegre *et al*. 1981,1982; Thompson *et al*. 1984; Zindler and Hart 1986). It is generally accepted that much of the asthenosphere is MORB-like and that OIB melts may be generated as small volume melts in the lithosphere, the asthenosphere, or in the deeper mantle. Detailed studies of the geochemistry of oceanic basalts (Zindler and Hart 1986) have produced evidence for several chemically distinct reservoirs: DMM or depleted MORB mantle, PREMA or prevalent mantle for ocean islands (e.g. Hawaii), HIMU or mantle with high μ (i.e.U/Pb), EM2 or enriched mantle 2, and EM1 or enriched mantle 1. An estimate of the composition of the bulk silicate earth from the work of Zindler and Hart (1986) compares favorably with that of Wanke (1981).

Within the continental regions the correlation between crustal age and kimberlite type helped define chemical provinces in the mantle beneath the Archaean Kaapvaal craton and the surrounding Proterozoic mobile belt in South Africa. Similarly, compilations of isotopic data for continental alkaline volcanic rocks from the western USA reveal isotopic provinciality (Leeman 1982). Menzies (1989) used xenolith-bearing volcanic rocks as a probe of the geometry and chemistry of lithospheric mantle domains and was able to assign a domain identity to subcrustal regions. In summary, the isotopic characteristics of the lithosphere beneath the mobile belts is similar to that beneath the ocean basins (Hart *et al*. 1989) and the lithosphere beneath stable cratons is to some degree isotopically unique. A close correlation exists between areas of high heat flow (>2 HFU), thin crust (<30km), and the eruption of continental volcanic rocks, whose source is in the asthenosphere, whereas continental volcanic rocks with a significant source contribution from the lithospheric mantle tend to be restricted to regions of low regional heat flow (<2 HFU) and thick crust (>40km). Chemical provinciality has also been recorded in the lithosphere beneath Scotland (Thirlwall 1982) and has similarly been interpreted as a result of lateral and vertical variability in the lithospheric mantle (Menzies and Halliday 1988). It should be noted that a shallow lithospheric source for continental alkaline basalts (Leeman 1982; Perry *et al*. 1987) from the western USA and Canada contrasts with the classic demonstration of an asthenospheric source for continental alkaline volcanic rocks from West Africa and the Gulf of Guinea (Fitton and Dunlop 1985). Indeed these data seem to indicate that under different circumstances alkaline melts can be generated in the convecting asthenosphere (or deeper) and in the lithosphere. If the chemical provinciality observed in continental volcanic rocks indicates the involvement of lithospheric mantle domains in their genesis. then continental volcanic rocks can be used to literally 'map' mantle domains (Fitton *et al*. 1988; Leat *et al*. 1988; Ormerod *et al*. 1988; Menzies 1989; Menzies and kyle, Chapter 8, this volume). Similarly, magma distribution and the variability in composition can be used to monitor the vertical movement of the asthenosphere-lithosphere boundary (McKenzie and Bickle 1988). It should be stressed that not all continental alkaline and tholeiitic volcanic rocks are suitable candidates for 'mapping' of the subcrustai lithosphere. Ewart *et al*. (1988) demonstrated that high-level processes involving continental crust may produce similar chemical provinciality in erupted basalts.

Continental flood basalts

The petrology of continental flood basalts was dealt with in many of the early texts (Washington 1922; Kuno 1969) and, more recently, Thompson (1977) deduced that certain flood basalt provinces were extruded at rates equivalent to mid-ocean ridge basalts. To account for these large volumes he proposed that continental basalts had an origin similar to oceanic basalts and that any modification of their trace element or isotopic ratio was due to digestion of variable quantities of crust during upwelling or accumulation of the magmas in the crust (e.g. Cox 1980). Whilst this is undoubtedly the case for several continental flood basalt provinces, including the British Tertiary, many others retain evidence of mantle heterogeneity. Leeman (1975,1977) and brooks and Hart (1978) proposed that the continental regions were underlain by lithospheric mantle that was very different from that beneath the ocean basins. Leeman (1977) believed in the existence of old enriched subcontinental mantle characterized by high Rb/Sr rtios and high $^{87}Sr/^{86}Sr$ rations and Brooks and Hart (1978) suggested that flood basalts inherited age information from their lithospheric mantle source, thus explaining the linear arrays of isotopic and trace element data. Hawkesworth *et al*. (1983,1986,1988) have more recently invoked the participation of Proterozoic lithospheric mantle in the production of the Karoo and the Parana flood basalt provinces. This was substantiated by recent Sm-Nd isotopic studies (Ellam and Cox 1989) which point to a Proterozoic age in the source region of some of the Karoo volcanic rocks. Pb isotope data for continental tholeiites from the western USA (Leeman 1975, 1977) also point to a Proterozoic rather than Archaean ages are forthcoming from these continental flood basalt provinces is of importance. The major, minor, and trace element chemistry of archaean lithoshpheric mantle is more refractory than post-Archaean lithospheric mantle which is closer to pyrolite in composition. It is therefore likely that such lherzolitic material could melt to produce vast quantities of tholeiitic magma than the refractory harzburgites found beneath the Archaean cratons. Whether continental flood basalts are derived from lithospheric mantle or asthenosphere with high-level contamination with crust remains unresolved in many cases (e.g. Hawkesworth and Norry 1983 and references therein; Moorbath *et al*. 1984 and references therein; MacDougall 1989 and references therences therein; Leeman and Fitton 1989 and references therein; Menzies and Kyle, Chapter 8, this volume).

Concluding statement

Volcanic rocks and xenoliths provide a unique insight into the petrology and geochemistry of the asthenosohere and lithosphere but it must be remembered that lherzolites and metasomatized xenoliths constitute a small proportion of the lithosphere. The continental lithospheric mantle may be largely comprised of a depleted protolith that remains in a ‘pristine’ state at some distance from magma conduits within the mechanical boundary layer and without the thermal boundary layer where the ingress of small volume melts from the asthenosphere produces a ‘secondary’ assemblage. The geochemistry of this depleted protolith holds the key to the evolution of the Archaean lithosphere, perhaps the only part of the upper mantle whose major element chemistry can be explained by extraction of crustal material (e.g. komatiites). Similarly, the precise significance of the lithosphere beneath Proterozoic and Phanerozoic terrains has not been thoroughly investigated. It can be demonstrated that Proterozoic and Phanerozoic lherzolites are a suitable source for tholeiitic melts with the isotopic characteristics of mid-ocean ridge basalts. The existence of such material beneath post-Archaean crust requires some explanation in that MORB mantle was depleted in the first billion years of Earth’s history. Moreover, the recent recognition of important correlations between seismic tomography and the geoid and tectonic features like subducted slabs may help focus our attention on active margin processes and the petrology and geochemistry of eclogites. Can active margin processes produce highly magnesian peridotites like those found beneath the Archaean crust? Finally, the importance of small volume melts and metasomatic processes are only now being realized with the study of wetting angles. The hitherto unspecified nature of metasomatic melts can now be more accurately defined (i.e. carbonate metls) but the exact relationship of such chemical transfer mechanisms to the origin of alkaline magmas is unknown as small volume melts can be generated within the asthenosphere and the lithosphere.

New words and expressions

peridotitic 橄榄岩的
Alkaline 碱性的
undepleted mantle 无亏损地幔
pyrolite 上幔岩，地幔岩
tholeiite(ic) 拉斑玄武岩（的）
polybaric 多压的
kimberlite 金伯利岩

volatile 挥发的
lherzolite 辉橄榄岩
domain 领域、区域，晶域，范围
matrix 母岩，基质
parent rock 母岩、原生岩
craton 克拉通, 即稳定地块
strontium 锶
micaceous 云母的,含云母的
basanite 碧玄岩，试金石
nephelinite
proterozoic 元古界
Archaean 太古代
metasomatic 交代的
conduit 管道、排水渠
geoid 地球体，地球形

spinel 尖晶石
eithophile 亲岩的，亲石的
diamond 金刚石
isotop 同位素
eclogite 榴辉岩
upwelling 上涌
harzburgite (saxonite) 方辉橄榄岩
protolith 母岩
komatiite 科马提岩
tomography 层面 X 线照相机

IX. Quaternary Geology

Pleistocene Nonmarine Environments

EDWARD S. EDDVEY, JR.[6]

Introduction

The chief early proponent of former glaciation was a biologist, and ever since Agassiz's day the role of biology in Pleistocene research has been one of full partner ship with physical geology. Other kinds of geology also depend heavily on biological data, but Pleistocene studies are primarily ecological, whereas the biology of older and longer time units is primarily evolutionary. The distinction is one of emphasis only, for the subjects are not sharply separable, but ecologists and evolutionists usually have different questions in mind. Similar or identical data—the occurrence of fossils in geologic settings—lead them to different inferences about the history of environments on the one hand, and about the phyletic history of organisms on the other.

It is true that organic changes that occurred during the Pleistocene are of exceptional interest to students of evolution, if only because the environmental setting of these changes is relatively well understood. The epoch was short, however, and the amount of morphologic and taxonomic change was slight, so that the leading problems are those of microsystematics rather than those of phylogeny. The evolutionist therefore can shift his attention from the deployment of organic variety to the mechanisms that produce and maintain the variety. However valuable its legacy to theory, the usefulness of this kind of evolution to stratigraphy and chronology is minimal. In other words, the evolutionary biologist expects to learn more from Pleistocene geology than he contributes to it.

By contrast, the history of environments is and must be inferred equally from physical and biological data. It is well known, even to meteorologists who have no concern with the geologic past, that vegetation is often a surer guide to the climate of a region, or to the microclimate of a locality, than any physical measurement yet devised. The central concern of Pleistocene research being the record of climatic variations, the ranges of animals and plants, when known to be environmentally determined and inferred to have changed from the distribution of fossils, are indispensable geologic documents. For the ecologists who provide and interpret these data it is a distinct advantage to be able to discount evolutionary change. It means that Pleistocene paleoecology deals for the most part with known or knowable species, and therefore proceeds in a more strictly uniformitarian way than is other kinds of paleoecology.

Pleistocene ecology is in fact nothing but an extension of present-day ecology into the most recent past. Major advances of ecological thought therefore tend to be reflected at once in Pleistocene studies, and the converse is also true. Because modern ecology is preoccupied with developing concepts of populations and ecosystems, a holistic

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viewpoint about life-in-environment is increasingly prevalent, and the metaphorical language of information theory is being usefully applied to systems, such as lakes and the ocean, that transmit biochemical order from the Pleistocene to the present. The systems approach is powerful, and it has the great merit that it appeals to the geochemists, whose recent contributions to paleoecology have been spectacular. As a less happy consequence, however, the autecologies of particular species of animals and plants are seen as trivial by comparison, and they are being less actively studied than formerly. Modern ecology is also undergoing a phase of intense self-criticism, and the current disenchantment with many long-established concepts, such as succession and the climax and ecogeographic rules like Bergmann's, is forcing a newly critical attitude in Pleistocene research. One result is that whereas two generations of workers have added huge quantities of information to a substantial base built by such workers as those of Hay (1923, 1924, 1927) and Baker (1920), the basis of ecological inference from a given fossil is often less certain now than it seemed in the golden days when there were fewer fossils to consider.

The function of this essay is not to give a comprehensive account of Pleistocene ecology, but to introduce and comment upon the contributions of biologists who review existing data on Pleistocene biogeography, including biological stratigraphy, in the United States. What emerges from these chapters, or at least from these comments, is a regrettable conclusion: extremely little is reliably known about Pleistocene habitats, especially at times and in places where there were no glaciers. Though it may surprise physical geologists, such a conclusion does not alarm general ecologists, who are aware of how little is reliably known about *any* habitat, ancient or modern. At all events, factual data are much more abundant than they were forty years ago, and the taxonomic quality of fossil identifications has been measurably improved. Skepticism in the geologic interpretation of these data is a healthy sign of scientific maturity, which the paleoecologist shares with the physical geologist.

Postglactal stratigraphy, Holocene

Postglacial pollen sequences are less well known in glaciated eastern North America than in Europe, the number of sections being a few hundred instead of several thousand, but the main outlines have been clear since the work of Paul B. Sears in Ohio in the 1930's. His five-part division as extended to Connecticut (Deevey, 1939, 1943) — A, boreal coniferous; B, pines; C-1, oak-hemlock; C-2, oak-hickory; and C-3, oak-chestnut-spruce-has been found to serve very well over most of this enormous region, correlations between zones being based on botanically appropriate modifications in regions where hemlock or hickory or chestnut are replaced as forest dominants by other tree species. It has been generally accepted that the implied climatic sequence —A, cool; B, warm, dry; C-1, warm, moister; C-2, warm, drier; C-3, cooler, moister—prevailed over an area much wider than the present range of any tree species; hence transatlantic and even worldwide eorelations have been attempted without diffidence, on the lines laid down by the master (von Post, 1946) in his Vega Lecture. It was also clear very early that a sequence beginning with a time of conifers must be all postglacial in the sense of Firbas and Iversen, and the search for older treeless or late-glacial pollen zones was successful (Deevey, 1951)as soon as the techniques of those workers were applied.

Radiocarbon dating was brought into use as early as 1950(Flint and Deevey, 1951; Deevey and Potzger, 1951) and verified the general synchrony of major pollen-analytical events on both sides of the Atlantic. Although the most direct test of correlation (comparison of Zone L-2 in Maine with the Allerod zone) required impracticably large C^{14} samples, it was possible to demonstrate equivalence of early Boreal time between Europe and southern New England or Michigan. Deevey and Flint (1957) therefore were able to discount the time-transgressive nature of the North American pine-pollen zone in defining the Hypsithermal interval (von Post's maximum-warmth period) as the time of Danish pollen-zones V through VIII.

Late-glacial

Several studies (Andersen, 1954; Leopold, 1956; Davis, 1958; Livingstone and Livingstone, 1958; Ogden, 1959; summary of daters by deevey, 1958) were concentrated on late glacial stratigraphy, in an effort to strengthen correlations by defining the zone of influence of the Valders glaciation. As it became clearer (Terasmae, 1959; MacClintock and Terasmae, 1960) that Valders ice probably did not occupy New England, unless as isolated remnant ice caps, southern New England became an obious place in which to search for evidence of pre-Allerod climatic oscillations.

By Two Creeks time southern New England and New York, like the type locality at Manitowoc, Wisconsin, but unlike Europe north of Spain, were evidently forested. Evidence of cooling beyond a periglacial belt was therefore to be sought, not in a classic tundra to park tundra to tundra sequence, but in some from of vegetational disturbance short of deforestation. In this situation it was essential to decide by what pollen-statistical criteria such disturbance could be recognized. European experience, with a different late-glacal geography and a floristically much simpler forest, provided no helpful parallels; moreover, the problem is not purely stratigraphic, but requires paleoecological interpretation of a kind that conventional pollen stratigraphy is poorly equipped to handle.

Statistical traps

Pollen percentages from a closed statistical universe, in which variation of one component affects numerical values for all other. Parallel trends of percentages over time at different sites define homotaxial zones, so accounting for the great success of the method as a stratigraphic tool; but as different plants vary enormously in production and dissemination of pollen, changing pollen percentages have no definable ecological meaning. This point was well understood by von Post, but his successors have tended to lose sight of it (Davis, 1963). It arose in acute from in connection with late-glacial variations of the ratio of spruce to non-arboreal pollen (Davis, 1961; Ogden, 1963), when inferences from pollen stratigraphy were sharply contradicted by radiocarbon dates at several sites on Martha's Vineyard and Block Island. Conifer pollen grains in a non-arboreal context are peculiarly liable to misinterpretation, because some or perhaps all of them have been blown for long distances over treeless country; but if the Totoket (Leopold, 1956) and Vineyard (Ogden, 1959) oscillations were statistical artifacts, other interpretations of changing percentages are equally questionable. For example, During postglacial warming (Zone B), when an aspen-birch woodland containing much and fir and then maple before the oaks arrive, one of the least obvious consequences is a rise of pine pollen from *ca*. 20% to *ca*. 60%.

The real meaning of the pine maximum is almost certainly *absence of oak,* but American palynologists have prevented from seeing this by the undoubted stratigraphic validity of the pine zone. Perhaps, too, they have been misled by the abundance of megafossils of *Pinus sylvestris* in European peat bogs, where the Boreal zone was defined before pollen analysis was invented. The “Sub-Atlantic climatic deterioration” (Zone C-3) is another problem demanding careful reconsideration of the meaning of pollen percentages. Davis (this volume; see also Smith, this volume) finds no evidence of this event, at least in New England pollen diagrams, that cannot be attributed with equal plausibility to the spread of agriculture in the first millennium B.C.

Uniformitarian approach

Late-Pleistocene paleoecology has the advantage of being able to work directly backward from the present. Martin and his co-workers in the Southwest, as is proper in any newly studied region, interpret buried pollen assemblages only in the light of the modern pollen rain in various habitats, and the lack of comparable information in the Northeast, though embarrassing, is repairable. Several studies (*e.g*. Carroll, 1943) have attempted to relate pollen rain to vegetation in quantitative terms, but moss and the uppermost surfaces of bogs are unsatisfactory samplers of modern pollen, because they represent indefinite times of accumulation. Glass slides and air filters work well for public-health purposes in cities, but their data are not easily translated into rates of natural pollen sedimentation. Analysis of the uppermost surfaces of lake deposits has become the method of choice, cattle tanks being used where natural lakes are not available, and although the limnology of pollen deposition (and redeposition) needs much more study, promising results are being obtained in several areas. Fortunately, as most pollen of paleoecological interest was waterlaid, it is sufficient for most purposes to measure pollen deposition on the lake bottom, so by passing serious aerodynamic problems posed by other kinds of fallout.

As present-day vegetation is reasonably well known in much of the Northeast, rapid reconnaissance in the lakes of several different regions gives useful results. Ogden (manuscript) has devised a surface-mud sampler for this purpose that is easily operated from a portable boat. So far as it relies on pollen percentages, however, the method is no different from Aario’s (1940), as applied to surface peat by Wilson, Hansen, and other American workers; it relates pollen dominance to plant cover in a non-quantitative way, but it is of little help in the difficult situation where a forest dominant, such as aspen or maple is virtually unrepresented by pollen. Dramatic examples of the lack of fit between pollen and vegetational frequencies are given by Davis and Goodlett (1960; so also Davis, 1963).

Pollen-fallout rates

To escape from the statistical straitjacket of percentage composition, chronological control is necessary. The fallout of each pollen type per unit area and time can be measured independently of other pollen types, if the sedimentation rate of unit volume or pollen-bearing deposit is known. Radiocarbon dating is the newest of several methods applicable to older deposits, and it has now yielded an absolute pollen diagram for the late-glacial history of Rogers Lake, Connecticut (Davis and Deevey, 1964). Current sedimentation rates in most lakes of cultivated regions differ sharply from those of a few centuries ago, however, and must be measured in many habitats if rations of pollen to vegetation are to be given ecological meaning. Some progress is being made in this direction, but the task is formidable, in view of the necessity of including arctic lakes.

Present-day sedimentation rates can be measured experimentally, in sediment traps, or, less directly, by dating short mud cores. Radiocarbon dating is too insensitive for this purpose, but other natural isotopes such as tritium and Pb^{210} offer promise. Annual laminae of the Zürichsee type (see discussion by Bradley, 1963) occur mainly in meromictic lakes, which are uncommon, but other stratigraphic markers that can be used include the incoming of European weed pollen, the post-1920 decline of chestnut (Castanea), and the post-1954 increase of artificial isotopes.

Direct measurement of pollen fallout in sediment traps suspended over the lake bottom has so far given ambiguous results in Rogers Lake (Davis and Deevey, 1962), mainly because of statistical uncertainties in counting. However subfossil *Bosmina* carapaces and organic matter in the same traps (Deevey, 1964)give mutually consistent figures averaging 3 years per milliliter (3.3 mm/yr)for the fresh sediment, compared to *ca*. 1mm/yr for the postglacial section (and 0.36 ± 0.33mm/yr for the dated late-glacial section; Davis and Deevey, 1964). If these flgures prove to be typical of small temperate lakes (Thomas, 1955; Tutin, 1955), *i.e*. if the variance of sedimentation rates is no greater than that of pollen counts, fallout rates may be adequately estimated from counts in standard volumes of surface sediment, without further refinement.

Absolute pollen frequency

Even in default of calibration between surface pollen frequencies and present-day vegetation, absolute pollen frequency (APF), or pollen per unit mass or volume, can be informative. On the assumption of constant deposition of dry matter, Tsukada (1958) has attempted to deduce relative pollen-dissemination rates of Japanese trees from changes of APF in different habitats. The practice of measuring and publishing APF values is to be encouraged, because they can be calibrated eventually, whereas percentages discard information that can not be recovered. When percentages and APF both change sharply, however, as they can be expected to do at the late-glacial-postglacial boundary, pollen frequencies remain equivocal as indicators of ecology.

Conclussion

Many other fossils besides pollen grains contribute to postglacial stratigraphy, and some, like small vertegrates in caves, terrestrial mollusks in loess, or seeds in swamps and salt marshes, are richly informative for paleoecology. Ross (this volume) emphasizes the high indicator value of insects, very many species of which have strict environmental requirements. This discussion has been confined to *pollen* stratigraphy because airborne pollen is nearly universal, and because a pattern of changing climax vegetation on uplands, inferable in principle from pollen, points more directly to climatic change than do events in lakes, swamps, or caves. To work out this pattern in quantitative terms has seemed eminently worth while to historians of all sorts, not least to archaeologists; and because of the primacy of plants in the environment of animals, including man, pollen stratigraphers have been able to think of themselves as custodians of the history of ecosystems. Other paleoecologists are aware that quantitative methods developed in postglacial stratigraphy are more or less directly applicable to countable assemblages of the remoter past, such as foraminiferas and diatoms in marine deposits as well as the pollen of lignite and the spores of coal. It turns out that the extreme ease with which pollen can be recovered, identified, and counted is one of nature's more seductive tricks. It has permitted a valid and useful stratigraphic division of late Pleistocene deposits, now largely superseded by radiocarbon dating, while the changing pattern of vegetation remains as elusive as ever.

The older Pleistocene, what happen to the Yarmouth ?

Pre-Wisconsin organic deposits are not common in North America, or, if they occur in non-glaciated regions, they are difficult to recognize. Sections that span a full interglacial age are extremely rare, and several of the better-known ones are not well placed stratigraphically. Of those studied pollen-analytically by Voss (1933,1939) and Fuller (1939), the majority have been reassigned to different stages. This is ordinarily done on glacial-geologic grounds, because Pleistocene pollen and other plant evidence lacks any built-in evolutionary clues to age, and interglacial sequences are inevitably much alike.

A part from the Aftonian beds of Iowa (Steere, 1942), well-studied eastern sections (Terasmae, 1960; Engelhardt, 1960; Kapp, 1964; Kapp and Gooding; 1964; Whitehead and Barghoorn, 1962) fall into two groups. Those dominated throughout by subarctic and boreal plants, implying a climate cooler than today's, are called interstadial. Only those with evidence of deciduous forest near the middle are considered to be interglacial, although this classic criterion breaks down in the far north (Terasmae, 1957). The problems of interpreting pollen percentages are no easier in truncated pre-or intra-Wisconsin section than in postglacial deposits, but the clear and consistent dominance of hardwood pollen in Sangamon diagrams from southern Indiana (Kapp and Gooding, 1964) leaves little doubt that that region had a rich hardwood forest, probably richer than today's, in middle-Sangamon time. In the famous Don beds near Toronto, Now also regarded as Sangamon, the presence of warmth-demanding plants was established on the basis of megafossils, and the pollen evidence is confirmatory. It may seem from the literature that the vegetation of the Northeast is better known in Sangamon than in Holocene time, but this is an illusion. Almost unconsciously, one expects ecological detail to be increasingly blurred in the remoter past, so that "hardwood forest," though acceptable as an index to interglacial climate, is much too imprecise for hypso-thermal time.

In the stratigraphic reassessment of many interglacial deposits, the Yarmouth interglacial appears to be a casualty. The concept of an American equivalent of the Great Interglacial is not to be dispensed with lightly, yet it is remarkable that no plant-bearing deposits can safely be called Yarmouth. The Puyallup interval in the Pacific Northwest (Crandell, this volume; see also Heusser, this volume) may be an equivalent, but there is no evidence from organic remains that this or nay other non-glacial interval was warmer than today in the Puget lowland. In fact, it is difficult to distinguish glacial from interglacial times in pollen assemblages from this region, where strong topographic contrasts and an unusual variety of wind-pollinated conifers contribute to the ambiguity. In the mid-continent region, south of the glacial boundaries, the deposits of Meade County, Kansas, have been studied by paleoecologists with particular care, yet Kapp (1964) describes no pollen bearing strata from this district older than Illinoian. As Taylor (this volume) knows no mollusk faunas of Yarmouth age, and as Hibbard *et al*. (this volume) can list only one mammal fauna (the Borchers) from this interval, a keystone of American Pleistocene stratigraphy is seen to be astonishingly loose.

A skeptical review of the data from the Great Plains, well summarized in chapters by Taylor, by Hibbard *et al*., and by Frye and Leonard, reinforces suspicion that the Yarmouth is a myth, or at least unrecognizable away from its type area, where it is mainly known as a profile of weathering. Evidence of alternating and erosion is abundant throughout the Plains, but the difficulties of correlating independent bodies of alluvium are notorious. Evolution having been too slow to be useful, even in mammals, ecological inferences have been used for correlation. The southern Plains were clearly better watered through much of the Pleistocene than they are today, and moisture demanding faunas are assigned to glacial stages while semiarid conditions more similar to the present are classed as interglacial. It is not necessary to challenge the well-established assumptions underlying this procedure, but the matching of these more and less pluvial episodes with particular stages in the glaciated districts depends on the further assumption that there were only four glacial and three interglacial ages, and the possible circularity is evident.

Faunal shifts on the great plains

Changes in rainfall in semiarid Kansas during the Pleistocene are made obvious by the periodic abundance of fishes, aquatic mammals and mollusks, and ostracods (for the last, wee Staplin, 1963). Such changes in a mid-continent region point to marked variation in degree of continentality, but not necessarily to net changes of temperature, and in fact there is little indication of climates much cooler or much warmer than today's in non-glaciated parts of the Plains. Taylor (this volume) points to the contrast in this respect between southwestern Kansas and such eastern regions as Indiana, and he notes that aridity becomes more important than temperature to animal ecology, in the Pleistocene as today, in the vicinity of the 100th meridian. In this situation the evidence from faunal shifts (Miller, this volume) is more equivocal than is generally realized.

For example, by mapping existing ranges of fossil fish species, and noting their area of overlap, Smith (1954 inferred that Kansas in Illinoian time was most like Wisconsin today, but although this was probably true in terms of fish habitats, the *climatic* similarity need not have been very close. More commonly, when assemblages of tens or hundreds of species are considered (Taylor, this volume), no exact modern counterpart of the fossil fauna can be found. By an interesting inversion of the principle of parsimony, the simplest explanation-random re-shuffling of ranges-is regarded by ecologists as the least likely. If only a few of the fossils seem out of context, and they are predominantly of northern distribution, one infers that the summer temperatures were cooler; conversely, southern species in otherwise uninformative assemblages imply that the winters were milder; and when extra limited species of both kinds occur together as fossils, the climate is thought to have been less continental. This reasoning is not inherently fallacious, but it assumes more knowledge of present-day ecology than can be obtained from maps. If species' present ranges were known to be at equilibrium, and the limits were known to be controlled by climate, latitudinal arrangements could be taken at face value. Unfortunately, even if one neglects such biotic factors as competition and sociality, it is possible for every segment of the limit of a mapped range to have a different control, as lacustrine species need standing water of the right chemistry, terrestrial animals need certain kinds of plants and/or trace minerals, and homoiothermal vertebrates have various ways of tempering the effects of climate.

Aware of all these difficulties, and others, Hibbard (1960)pictures Pleistocene climates in the southern Plains as not very different from today's, except for the periodically greater rainfall, and he suggests that glacial-interglacial contrasts were notably less sharp beyond the drift borders than is implied by the glacial sequence itself. Like Taylor (this volume), Hibbard considers the Wisconsin as the time of maximum climatic stress, so accounting for the relatively abrupt extinction of many animals, and particularly of such thermophiles as the tortoise *Geochelone*, that had persisted through earlier cycles. Although this view seems to have been arrived at rather reluctantly, it is consistent with the botanical data from the Southwest(Martin, this volume), where spruce pollen and other indications of cool climate are strangely lacking in pre-Wisconsin levels.

Between Meade County, Kansas, where Hibbard and Taylor have worked and the San Augustin Plains, New Mexico, the site of the deep boring of Clisby and Sears (1956), two pre-Wisconsin lake deposits in Oklahoma and in Texas have been studied with great skill by Stephens (1960) and by Kirkland and Anderson (1963). The lacustrine sequences are short, of different ages, and dated only roughly as Late and Early Pleistocene by their few mammalian fossils and their relation to the Pearlette ash. Both imply pluvial conditions, but the only indications of cooling are those that would be expected to accompany increased moisture. Despite the discontinuities in the record from the southern Plains, it begins to seem that this prairie and semi-desert region was at most a warm savanna, never a cold steppe, throughout the Pleistocene.

Plio-Pleistocene boundary

As in the interpretation of the late-glacial in the Northeast, American paleoecologists have been influenced, and possibly misled, by the Pleistocene of central and southeastern Europe. It is easy for Americans to underrate the significance of Europe's transverse mountain system, and of the access it gives, even today, to a cold-steppe fauua and flora from Asia. A natural supposition that American glacial ages ought to be as cold as Europe's, combined with an *a priori* belief in four glaciations, with Nebraskan equal to Günz, may have encouraged correlations between Kansas and Illinois that have little factual basis. If so, the Yarmouth interglacial may not be the only casualty of the coming stratigraphic reassessment; the Aftonian of the Plains also needs to be re-examined.

The scanty vertebrate faunas called Nebraskan and Aftonian are also Blancan, and they therefore may be of Villafranchian age if not older (Evernden *et al*., 1964). The case for a pre-Günz Pleistocene in Europe does not rest on the problematical Donau glaciation, but on clear faunistic and floristic evidence that the Villafranchian fauna is older than the Cromerian interglacial. If there are American equivalents of the Pre-Cromerian, early-Pleistocene cold phases of the Netherlands (Zagwijn, 1960), they are probably to be sought above the "Aftonian" of Kansas, though the type Aftonian beds of Iowa may still be Cromerian and the Nebraskan still correlative with Günz.

Long persistence of an "archaic" or Pliocene fauna, once thought to have been an African peculiarity, is now rather generally accepted for regions that never experienced continental glaciations; western North America is not atypical in having the Blancan faunal zone, Pleistocene only in its upper part. It need not follow from this that the Nebraskan glaciation is Middle Pleistocene, but the possibility can no longer be discounted. Among other attractions, it would harmonize the shorter and the current longer estimates of the duration of the Pleistocene by dividing the epoch into glacial and nongacial portions of nearly equal length. Moreover, despite the way in which the classical stage-names are used in sea-floor stratigraphy (Ericson *et al*., 1961, 1963, 1964), the suggestion appears not to conflict with the data of that field. Correlation of cold and warm foraminifera assemblages, particularly the older ones, is hampered by unconformities and reduplications and the paucity of stratigraphic markers_ a situation reminiscent of alluvial stratigraphy on the Plains-and the assignment of these marine faunas to continental stages, by the Lamont group as by Emiliani (1955), seems to reflect standard nomenclatural preconceptions.

If parts of the extraglacial American Pleistocene are older than has been believed, as well as more similar climatically to the present, similarity to the upper Pliocene would not be remarkable, before extensive continental glaciation occurred, neither the range of environments nor the stage of mammalian evolution should show any major discontinuity. The Plio-Pleistocene boundary is as inconspicuous in the long pollen sequence beneath the San Augustin Plains (Clisby and Sears, 1956) as it is in the Blancan faunal zone.

The apparent complacency of the early record in this core, and the increase of spruce pollen in and only in the Wisconsin levels, have been attributed to late-Pleistocene uplift of the basin (Clisby and foreman, 1958). Although late-Tertiary local and regional upwarming of the Cordillera is generally accepted and is commonly invoked to account for increased continentality and regional climatic differentiation toward the east, there is no independent evidence that New Mexico was uplifted so recently by the magnitude required (several thousand feet) to bring its summits from the juniper to the spruce belt. Moreover, the Holocene diminution of spruce pollen, well established thoughtout the Southwest, is unlikely to have been the result of crustal subsidence.

Wherever else it lies, the Plio-Pleistocene boundary is somewhere between the bottom of the glens Ferry formation in Idaho (Malde, this volume) and the bottom of the overlying Bruneau beds, *i.e*. between potassium-argon dates of 3.5 and 1.4×10^6 yr, and there is good reason to expect it within the exposed 2,000 ft of the fabulous Glenns Ferry. When studies of the pollen, by Estella Leopold, and of the mollusks, by Dwight Taylor, are complete, the boundary will probably turn out to be as obscure as it is in the San Augustin core.

If these tentative, unorthodox views of Pleistocene ecology and stratigraphy are verified, delayed onset of continental glaciation and its apparent culmination in the Wisconsin age will demand explanation, and the explanations may not be the same. Late-Tertiary uplift may account for the first, while the oceanography of the Arctic basin may hold the key to the second, as Taylor (this volume) suggests. Studies of the submarine stratigraphy of that basin are awaited with great interest.

New words and expressions

Pleistocene 更新世
chronology 年代学，地质年代、年表
rolistic 整体的
microsyste matics 微观系统学
ecogeography 生态地理学
methodology 方法学方法论
pluvial 洪水的, 多雨的
boreo-alpine 北方高山的
hydrology 水文学
carnivores 食肉动物,食虫植物
indigenous 土著的，乡土的
hypsithermal （冰后期）高温的
savanna 萨瓦纳群落，热带稀树干草原
pollen 花粉
oak-hemlock 栎-铁杉
spruce 云杉
nonarboreal 非林的，树木的
fir 冷杉
maple 枫树、槭树
moss 苔藓，地衣，泥类沼泽
limnology 湖沼学
lignite 褐煤
deciduous 落叶的
lacustrine 湖泊的
thermophile 喜温生物
physical geology 普通地质
meteorologist 气象学家
phylogeny 系统发生、种系发生
autecology 个体生态学
biogeography 生物地理学
niche 小生境
sympatric（生态）分布区重叠的
loess 黄土
endemic 土著的，本地的
herbivose 食草动物
biota 生物群
steppe 草原
pines 松树
hickory 山核桃
tundra 冻原，苔原
aspen-birch 杨-桦
larch 落叶松
palynologist 孢粉学家
bog 沼泽，泥塘
archaeologist 考古学家
infersfadial 间冰期
alluvion 冲积层
homoisothermal 同等温的
prairie 大草原