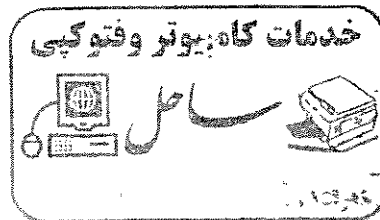University of Birjand
Faculty of Sciences

Scientific texts in Geology

(Geology in English)

Complied by: M. Motamedalshariati

2020

Minerals: Building Blocks of Rocks

"Everything should be made as simple as possible, but no simpler."

ALBERT EINSTEIN

In Chapter 2, we saw how plate tectonics describes Earth's large-scale structure and dynamics, but we touched only briefly on the wide variety of materials that appear in plate tectonic settings. In this chapter and the next, we focus on rocks, the records of geologic processes, and on minerals, the building blocks of rocks.

Rocks and minerals help determine various parts of the Earth system, much as concrete, steel, and plastic determine the structure, design, and architecture of large buildings. To tell Earth's story accurately, geologists often adopt a "Sherlock Holmes" approach: they use current evidence to deduce the processes and events that occurred in the past at some particular place. The kinds of minerals found in volcanic rocks, for example, give evidence of eruptions that brought molten rock, at temperatures perhaps as high as 1000°C, to Earth's surface. The minerals of a granite reveal that it crystallized deep in the crust under the very high temperatures and pressures that occur when two continental plates collide and form mountains like the Himalayas. Understanding the geology of a region allows us to make informed guesses about where undiscovered deposits of economically important mineral resources might lie.

We turn now to the focus of this chapter: **mineralogy**—the branch of geology that studies the composition, structure, appearance, stability, occurrence, and associations of minerals.

What Are Minerals?

Minerals are the building blocks of rocks: with the proper tools, most rocks can be separated into their constituent minerals. A few kinds of rocks, such as limestone,
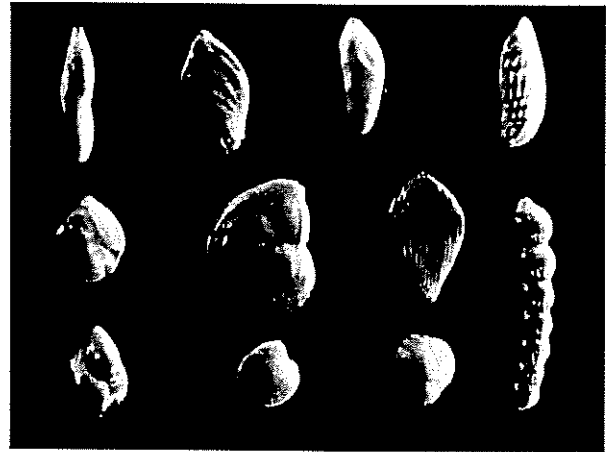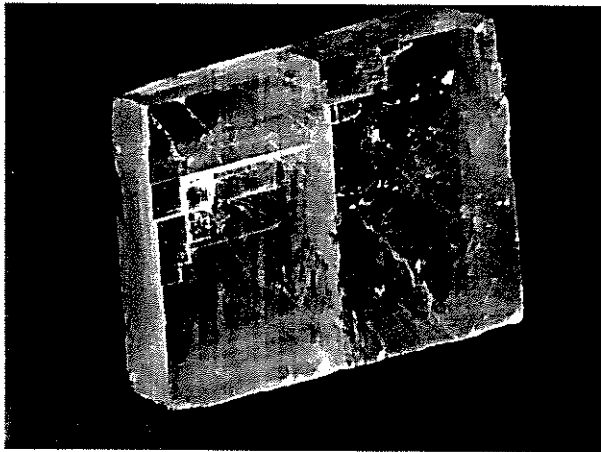
Figure 3.1 The mineral calcite is found in the shells of many organisms, such as foraminifera.
[*left:* Lester V. Bergman/Corbis; *right:* Cushman Foundation for Foraminiferal Research, 1987.]

contain only a single mineral (in this case, calcite). Other rocks, such as granite, are made of several different minerals. To identify and classify the many kinds of rocks that compose the Earth and understand how they formed, we must know about minerals.

Geologists define a **mineral** as a *naturally occurring, solid crystalline substance, generally inorganic, with a specific chemical composition.* Minerals are homogeneous: they cannot be divided by mechanical means into smaller components.

Let us examine each part of our definition of a mineral in a little more detail.

Naturally Occurring ... To qualify as a mineral, a substance must be found in nature. Diamonds mined from the Earth in South Africa are minerals. Synthetic versions produced in industrial laboratories are not considered to be minerals. Nor are the thousands of laboratory products invented by chemists.

Solid Crystalline Substance ... Minerals are solid substances—they are neither liquids nor gases. When we say that a mineral is *crystalline,* we mean that the tiny particles of matter, or atoms, that compose it are arranged in an orderly, repeating, three-dimensional array. Solid materials that have no such orderly arrangement are referred to as *glassy* or *amorphous* (without form) and are not conventionally referred to as minerals. Windowpane glass is amorphous, as are some natural glasses formed during volcanic eruptions. Later in this chapter, we will explore in detail the process by which crystalline materials form.

Generally Inorganic ... Minerals are defined as inorganic substances and so exclude the organic materials that make up plant and animal bodies. Organic matter is composed of organic carbon, the form of carbon found in all organisms, living or dead. Decaying vegetation in a swamp may be geologically transformed into coal, which also is made of organic carbon; but although it is found as a natural deposit, coal is not traditionally considered a mineral. Many minerals are, however, secreted by organisms. One such mineral, calcite (**Figure 3.1**), forms the shells of oysters and many other organisms, and it contains inorganic carbon. The calcite of these shells, which constitute the bulk of many limestones, fits the definition of a mineral because it is inorganic and crystalline.

... With a Specific Chemical Composition The key to understanding the composition of Earth's materials lies in knowing how the chemical elements are organized into minerals. What makes each mineral unique is the combination of its chemical composition and the arrangement of its atoms in an internal structure. A mineral's chemical composition either is fixed or varies within defined limits. The mineral quartz, for example, has a fixed ratio of two atoms of oxygen to one atom of silicon. This ratio never varies, although quartz is found in many different kinds of rock. The components of the mineral olivine—iron, magnesium, and silicon—always have a fixed ratio. Although the ratio of iron to magnesium atoms may vary, the sum of those atoms in relation to the number of silicon atoms always forms a fixed ratio.

The Atomic Structure of Matter

A modern dictionary lists many meanings for the word *atom* and its derivatives. One of the first is "anything considered the smallest possible unit of any material." To the ancient

Greeks, *atomos* meant "indivisible." John Dalton (1766–1844), an English chemist and the father of modern atomic theory, proposed that atoms are particles of matter of several kinds that are so small that they cannot be seen with any microscope and so universal that they compose all substances. In 1805, Dalton hypothesized that each of the various chemical elements consists of a different kind of atom, that all atoms of any given element are identical, and that chemical compounds are formed by various combinations of atoms of different elements in definite proportions.

By the early twentieth century, physicists, chemists, and mineralogists, building on Dalton's ideas, had come to understand the structure of matter much as we do today. We now know that an **atom** is the smallest unit of an element that retains the physical and chemical properties of that element. We also know that atoms are the small units of matter that combine in chemical reactions and that atoms themselves are divisible into even smaller units.

The Structure of Atoms

Understanding the structure of atoms allows us to predict how chemical elements will react with one another and form new crystal structures. For more detailed information about the structure of atoms, see Appendix 4.

The Nucleus: Protons and Neutrons At the center of every atom is a dense **nucleus** containing virtually all the mass of the atom in two kinds of particles: protons and neutrons (Figure 3.2). A **proton** has a positive electrical charge of +1. A **neutron** is electrically neutral—that is, uncharged. Atoms of the same chemical element may have different numbers of neutrons, but the number of their protons does not vary. For instance, all carbon atoms have six protons.

Electrons Surrounding the nucleus is a cloud of moving particles called **electrons,** each with a mass so small that it is conventionally taken to be zero. Each electron carries an electrical charge of −1. The number of protons in the nucleus of any atom is balanced by the same number of electrons in the cloud surrounding the nucleus, so an atom is electrically neutral. Thus the nucleus of the carbon atom is surrounded by six electrons (see Figure 3.2).

Atomic Number and Atomic Mass

The number of protons in the nucleus of an atom is called its **atomic number.** Because all atoms of the same element have the same number of protons, they also have the same atomic number. All atoms with six protons, for example, are carbon atoms (atomic number 6). In fact, the atomic number of an element can tell us so much about an element's behavior that the periodic table organizes elements according to their atomic number (Figure 3.3). Elements in the same vertical group, such as carbon and silicon, tend to react similarly. For more detail about the periodic table, see Appendix 4.
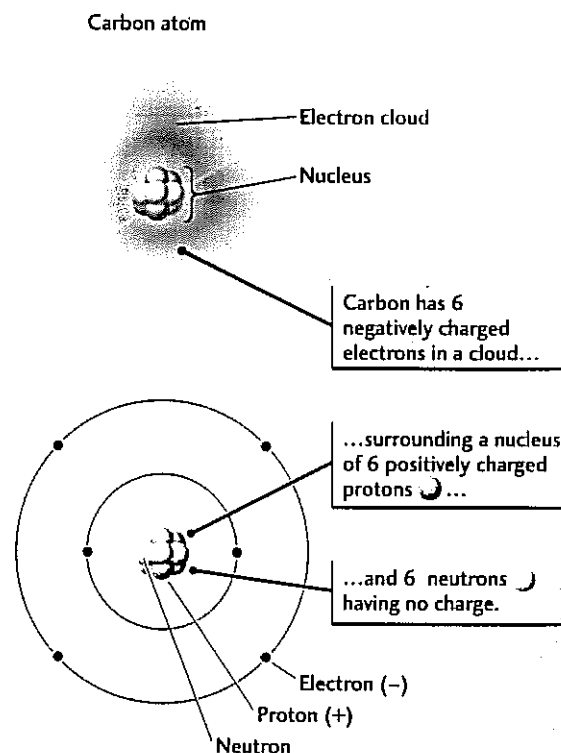
Carbon atom

Figure 3.2 Electron structure of the carbon atom (carbon-12). The electrons, each with a charge of −1, are represented as a negatively charged cloud surrounding the nucleus, which contains six protons, each with a charge of +1, and six neutrons, each with zero charge. The size of the nucleus is greatly exaggerated; it is much too small to show at a true scale.

The **atomic mass** of an element is the sum of the masses of its protons and neutrons. (Electrons, because they have so little mass, are not included in this sum.) Although the number of protons is constant, atoms of the same chemical element may have different numbers of neutrons and therefore different atomic masses. These various kinds of atoms are called **isotopes.** Isotopes of the element carbon, for example, all with six protons, exist with six, seven, and eight neutrons, giving atomic masses of 12, 13, and 14.

In nature, the chemical elements exist as mixtures of isotopes, so their atomic masses are never whole numbers. Carbon's atomic mass, for example, is 12.011. It is close to 12 because the isotope carbon-12 is overwhelmingly abundant. The relative abundance of the different isotopes of an element on Earth is determined by processes that enhance the abundance of some isotopes over others. Carbon-12, for example, is favored by some reactions, such as photosynthesis, in which organic carbon compounds are produced from inorganic carbon compounds.

Physical Properties of Minerals

Geologists use their knowledge of mineral composition and structure to understand the origins of rocks. First, they must identify the minerals that compose a rock. To do so, they rely greatly on chemical and physical properties that can be observed relatively easily. In the nineteenth and early twentieth centuries, geologists carried field kits for the rough chemical analysis of minerals that would help in identification. One such test is the origin of the phrase "the acid test." It consists of dropping diluted hydrochloric acid (HCl) on a mineral to see if it "fizzes" (**Figure 3.16**). The fizzing indicates that carbon dioxide (CO_2) is escaping, which means that the mineral is likely to be calcite, a carbonate mineral.

In the rest of this chapter, we will review the physical properties of minerals, many of which contribute to their practical and decorative value (see Feature 3.1).

Hardness

Hardness is a measure of the ease with which the surface of a mineral can be scratched. Just as a diamond, the hardest mineral known, scratches glass, so a quartz crystal, which is harder than feldspar, scratches a feldspar crystal. In 1822, Friedrich Mohs, an Austrian mineralogist, devised a scale (now known as the **Mohs scale of hardness**) based on the ability of one mineral to scratch another. At one extreme is the softest mineral (talc); at the other, the hardest (diamond) (Table 3.2). The Mohs scale is still one of the best practical tools for identifying an unknown mineral. With a knife blade and a few of the minerals on the hardness scale, a field geologist can gauge an unknown mineral's position

on the scale. If the unknown mineral is scratched by a piece of quartz but not by the knife, for example, it lies between 5 and 7 on the scale.

Recall that covalent bonds are generally stronger than ionic bonds. The hardness of any mineral depends on the strength of its chemical bonds: the stronger the bonds, the harder the mineral. Crystal structure varies in the silicate group of minerals, and so does hardness. For example, hardness varies from 1 in talc, a sheet silicate, to 8 in topaz, a silicate with isolated tetrahedra. Most silicates fall in the 5 to 7 range on the Mohs scale. Only sheet silicates are relatively soft, with hardnesses between 1 and 3.

Table 3.2 Mohs Scale of Hardness

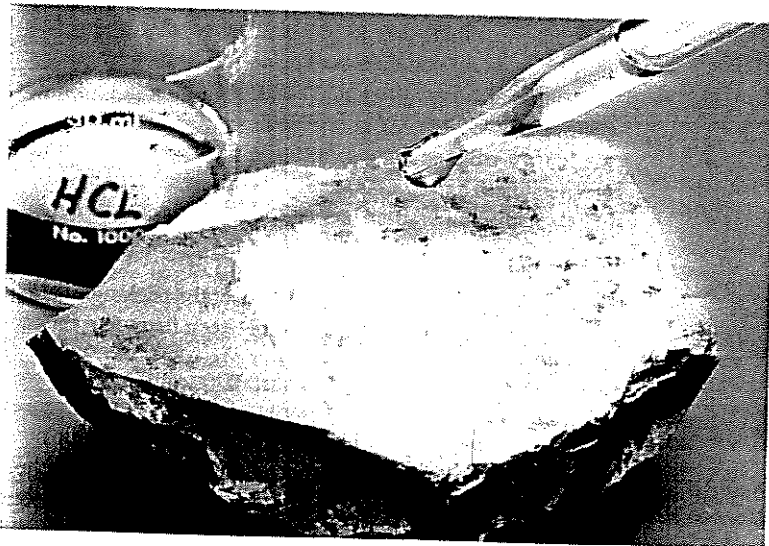| Mineral	Scale Number	Common Objects
Talc	1	
Gypsum	2	—— Fingernail
Calcite	3	—— Copper coin
Fluorite	4	
Apatite	5	—— Knife blade
Orthoclase	6	—— Window glass
Quartz	7	—— Steel file
Topaz	8	
Corundum	9	
Diamond	10	

Figure 3.16 The acid test. One easy but effective way to identify certain minerals is to drop diluted hydrochloric acid (HCl) on the substance. If it fizzes, indicating the escape of carbon dioxide, the mineral is likely to be calcite. [Chip Clark.]

EARTH ISSUES

## 3.1 What Makes Gems So Special?

No one can be sure when the first human picked up a mineral crystal and kept it for its rare beauty, but we do know that gems were being worn as necklaces and other adornments at the dawn of civilization in Egypt, at least 4000 years ago. These early Egyptians were undoubtedly attracted to the color and play of light on the polished surfaces of such minerals as carnelian, lapis lazuli, and turquoise. Color and luster (the ability to reflect light) are two qualities that still serve to define gemstones. Although the value placed on a gemstone varies in different cultures and historical periods, other required qualities seem to be beauty, transparency, brilliance, durability, and rarity. Most minerals have some of these remarkable qualities, but the stones considered most precious are ruby; sapphire; emerald; and, of course, diamond.

A diamond—geologically speaking, at least—may not be forever, as the advertisers claim, but it is special. Its glitter is unique, as are the play of colors and the sparkle that it emits. The source of these qualities is the way in which diamond refracts, or bends, light. These characteristics are enhanced by diamond's remarkable ability to split perfectly along certain directions of the crystal, which diamond cutters use to advantage in carefully cutting gem-quality stones. Diamond's multiple facets (faces superficially similar to crystal faces) can be polished to enhance this sparkle. These facets can be ground only by other diamonds, because diamond is the hardest mineral known—so hard that it can scratch any other mineral and remain undamaged. This mineral's tightly packed crystal structure and strong covalent bonds between carbon atoms give it these characteristics, which allow it to be identified with certainty by mineralogists and jewelers.

Rubies and sapphires are gem-quality varieties of the common mineral corundum (aluminum oxide), which is widespread and abundant in a number of rock types. Although not as hard as diamond, corundum is extremely

hard. Small amounts of impurities produce the intense colors that we value. Ruby, for example, is red because of small amounts of chromium, the same substance that gives emeralds their green color.

Less valuable, sometimes called semiprecious, gemstones are topaz, garnet, tourmaline, jade, turquoise, and zircon. Most, like garnet, are common constituents of rocks, occurring mostly as small imperfect crystals with many impurities and poor transparency. But, under special conditions, gem-quality garnets form. From time to time, some minerals that are not ordinarily considered gems may enjoy sudden—perhaps temporary—popularity. Hematite (iron oxide) currently enjoys this status, appearing in necklaces and bracelets.
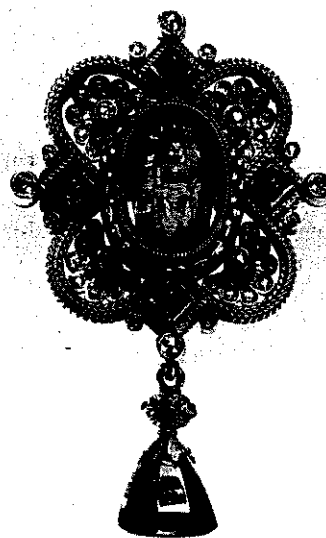

Sapphire (blue) and diamond (colorless) brooch by Fortunato Pio Castellani, Smithsonian Institution, nineteenth century. [Aldo Tutino/Art Resource.]

Within groups of minerals having similar crystal structures, increasing hardness is related to other factors that also increase bond strength:

• *Size:* The smaller the atoms or ions, the smaller the distance between them, the greater the electrical attraction, and thus the stronger the bond.

• *Charge:* The larger the charge of ions, the greater the attraction between ions and thus the stronger the bond.

• *Packing of atoms or ions:* The closer the packing of atoms or ions, the smaller the distance between them and thus the stronger the bond.

Size is an especially important factor for most metallic oxides and for most sulfides of metals with high atomic numbers—such as gold, silver, copper, and lead. Minerals of these groups are soft, with hardnesses of less than 3, because their metallic cations are so large. Carbonates and sulfates, groups
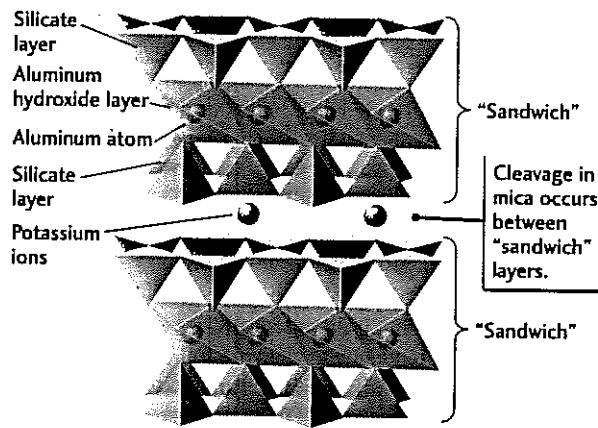
Figure 3.17 Cleavage of mica. The diagram shows the cleavage planes in the mineral structure, oriented perpendicular to the plane of the page. Horizontal lines mark the interfaces of silica-oxygen tetrahedral sheets and sheets of aluminum hydroxide bonding the two tetrahedral layers into a "sandwich." Cleavage takes place between composite tetrahedral–aluminum hydroxide sandwiches. The photograph shows thin sheets separating along the cleavage planes. [Chip Clark.]

in which the structures are packed less densely, also are soft, with hardnesses of less than 5. In all these groups, the hardness reflects the strength of the chemical bonds.

Cleavage

Cleavage is the tendency of a crystal to break along flat planar surfaces. The term is also used to describe the geometric pattern produced by such breakage. Cleavage varies inversely with bond strength: high bond strength produces poor cleavage; low bond strength produces good cleavage. Because of their strength, covalent bonds generally give poor or no cleavage. Ionic bonds are relatively weak, so they give excellent cleavage.

If the bonds between some of the planes of atoms or ions in a crystal are weak, the mineral can be made to split along those planes. Muscovite, a mica sheet silicate, breaks along smooth, lustrous, flat, parallel surfaces, forming transparent sheets less than a millimeter thick. Mica's excellent cleavage results from weakness of the bonds between the sandwiched layers of cations and tetrahedral silica sheets (Figure 3.17).

Cleavage is classified according to two primary sets of characteristics: (1) the number of planes and pattern of cleavage, and (2) the quality of surfaces and ease of cleaving.

Number of Planes; Pattern of Cleavage The number of planes and patterns of cleavage are identifying hallmarks of many rock-forming minerals. Muscovite, for example, has only one plane of cleavage, whereas calcite and dolomite have three excellent cleavage directions that give them a rhomboidal shape (**Figure 3.18**).

Figure 3.18 Example of rhomboidal cleavage in calcite. Calcite can be cleaved by a light hammer blow on a chisel oriented parallel to one of its planes. [Chip Clark.]

A crystal's structure determines its cleavage planes and its crystal faces. Crystals have fewer cleavage planes than possible crystal faces. Faces may be formed along any of numerous planes defined by rows of atoms or ions. Cleavage occurs along any of those planes across which the bonding is weak. All crystals of a mineral exhibit its characteristic cleavage, whereas only some crystals display particular faces.

Galena (lead sulfide, PbS) and halite (sodium chloride, NaCl) cleave along three planes, forming perfect cubes. Distinctive angles of cleavage help identify two important groups of silicates, the pyroxenes and amphiboles, that otherwise often look alike (**Figure 3.19**). Pyroxenes have a single-chain linkage and are bonded so that their cleavage planes are almost at right angles (about 90°) to each other. In cross section, the cleavage pattern of pyroxene is nearly a square. In contrast, amphiboles, the double chains, bond to give two cleavage planes, at about 60° and 120° to each other. They produce a diamond-shaped cross section.

Quality of Surfaces; Ease of Cleaving A mineral's cleavage is assessed as perfect, good, or fair, according to the quality of surfaces produced and the ease of cleaving. Muscovite can be cleaved easily, producing extremely high quality, smooth surfaces; its cleavage is *perfect*. The single- and double-chain silicates (pyroxenes and amphiboles, respectively) show *good* cleavage. Although these minerals break easily along the cleavage plane, they also break across it, producing cleavage surfaces that are not as smooth as those of mica. *Fair* cleavage is shown by the ring silicate beryl. Beryl's cleavage is less regular, and the mineral breaks relatively easily along directions other than cleavage planes.

Many minerals are so strongly bonded that they lack even fair cleavage. Quartz, a framework silicate, is so strongly bonded in all directions that it breaks only along irregular surfaces. Garnet, an isolated tetrahedral silicate, also is bonded strongly in all directions and so has no cleavage. This absence of a tendency to cleave is found in most framework silicates and in silicates with isolated tetrahedra.
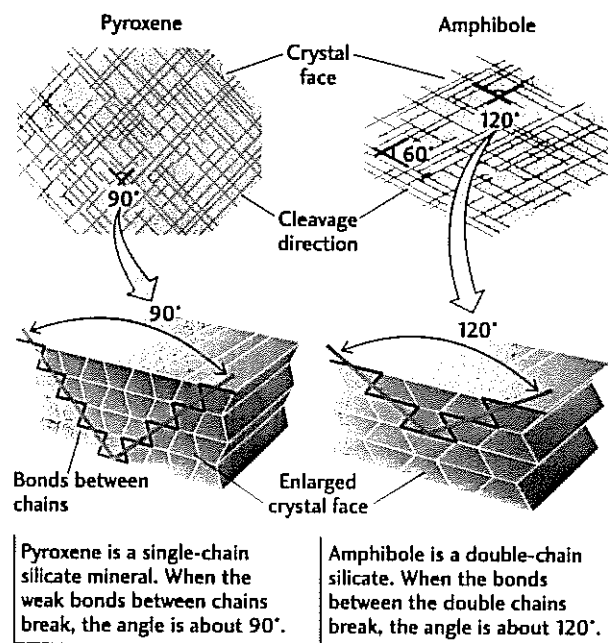
Fracture

Fracture is the tendency of a crystal to break along irregular surfaces other than cleavage planes. All minerals show fracture, either across cleavage planes or—in minerals such as quartz, with no cleavage—in any direction. Fracture is related to how bond strengths are distributed in directions that cut across crystal planes. Breakage of these bonds results in irregular fractures. Fractures may be *conchoidal,* showing smooth, curved surfaces like those of a thick piece of broken glass. A common fracture surface with an appearance like split wood is described as *fibrous* or *splintery.* The shape and appearance of many kinds of irregular fractures depend on the particular structure and composition of the mineral.

Luster

The way in which the surface of a mineral reflects light gives it a characteristic **luster.** Mineral lusters are described by the terms listed in Table 3.3. Luster is controlled by the kinds of atoms present and their bonding, both of which affect the way light passes through or is reflected by the mineral. Ionically bonded crystals tend to be glassy, or vitreous, but covalently bonded materials are more variable. Many tend to have an adamantine luster, like that of diamond. Metallic luster is shown by pure metals, such as gold, and by many sulfides, such as galena (lead sulfide, PbS). Pearly luster results from multiple reflections of light from planes beneath the surfaces of translucent minerals, such as the mother-of-pearl inner surfaces of many clam shells, which are made of the mineral aragonite. Luster, although an important criterion for field classification, depends heavily on the visual perception of reflected light. Textbook descriptions fall short of the actual experience of holding the mineral in your hand.

Color

The **color** of a mineral is imparted by light—either transmitted through or reflected by crystals, irregular masses, or a streak. **Streak** refers to the color of the fine deposit of mineral dust left on an abrasive surface, such as a tile of

Pyroxene   Amphibole

Crystal face

120°

60°

90°

Cleavage direction

90°

120°

Bonds between chains

Enlarged crystal face

| Pyroxene is a single-chain silicate mineral. When the weak bonds between chains break, the angle is about 90°. | Amphibole is a double-chain silicate. When the bonds between the double chains break, the angle is about 120°. |

Figure 3.19 Comparison of cleavage directions and typical crystal faces in pyroxene and amphibole. These two minerals often look very much alike, but their angles of cleavage differ. These angles are frequently used to identify and classify them.

Table 3.3 Mineral Luster

Luster	Characteristics
Metallic	Strong reflections produced by opaque substances
Vitreous	Bright, as in glass
Resinous	Characteristic of resins, such as amber
Greasy	The appearance of being coated with an oily substance
Pearly	The whitish iridescence of such materials as pearl
Silky	The sheen of fibrous materials such as silk
Adamantine	The brilliant luster of diamond and similar minerals

unglazed porcelain, when a mineral is scraped across it. Such tiles are called *streak plates* (Figure 3.20). A streak plate is a good diagnostic tool because the uniform small grains of mineral that are present in the powder on the ceramic tile permit a better analysis of color than does a mass of the mineral. A mass formed of hematite (Fe_2O_3), for example, may be black, red, or brown, but this mineral will always leave a trail of reddish brown dust on a streak plate.
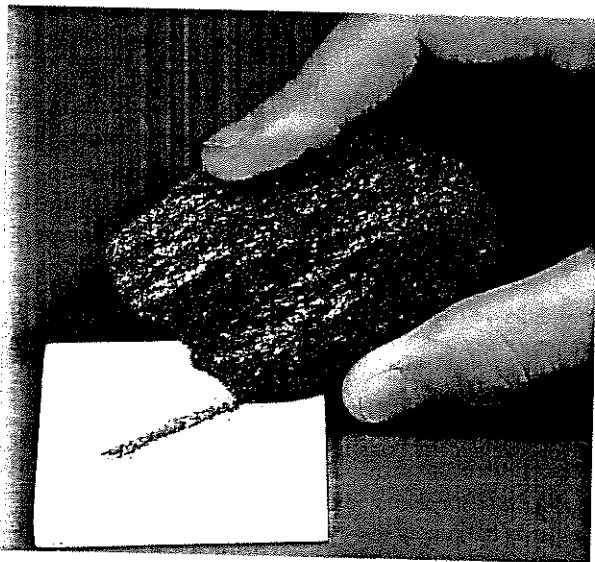
Figure 3.20 Hematite may be black, red, or brown, but it always leaves a reddish brown streak when scratched along a ceramic plate. [Breck P. Kent.]

Color is a complex and not yet fully understood property of minerals. It is determined both by the kinds of ions found in the pure mineral and by trace impurities.

Ions and Mineral Color The color of pure substances depends on the presence of certain ions, such as iron or chromium, that strongly absorb portions of the light spectrum. Olivine that contains iron, for example, absorbs all colors except green, which it reflects, so we see this type of olivine as green. We see pure magnesium olivine as white (transparent and colorless).

Trace Impurities and Mineral Color All minerals contain impurities. Instruments can now measure even very small quantities of some elements—as little as a billionth of a gram in some cases. Elements that make up much less than 0.1 percent of a mineral are reported as "traces," and many of them are called trace elements.

Some trace elements can be used to interpret the origins of the minerals in which they are found. Others, such as the trace amounts of uranium in some granites, contribute to local natural radioactivity. Still others, such as small dispersed flakes of hematite that color a feldspar crystal brownish or reddish, are notable because they give a general color to an otherwise colorless mineral. Many of the gem varieties of minerals, such as emerald (green beryl) and sapphire (blue corundum), get their color from trace impurities dissolved in the solid crystal (see Feature 3.1). Emerald derives its color from chromium; the source of sapphire's blue color is iron and titanium.

The color of a mineral may be distinctive, but it is not the most reliable clue to its identity. Some minerals always show the same color; others may have a range of colors. Many minerals show a characteristic color only on freshly broken surfaces or only on weathered surfaces. Some—precious opals, for example—show a stunning display of colors on reflecting surfaces. Others change color slightly with a change in the angle of the light shining on their surfaces.

Specific Gravity and Density

One can easily feel the difference in weight between a piece of hematite iron ore and a piece of sulfur of the same size by hefting the two pieces. A great many common rock-forming minerals, however, are too similar in density—mass per unit volume (usually expressed in grams per cubic centimeter, g/cm^3)—for such a simple test. Scientists therefore need some easy method for measuring this property of minerals. A standard measure of density is **specific gravity**, which is the weight of a mineral in air divided by the weight of an equal volume of pure water at 4°C.

Density depends on the atomic mass of a mineral's ions and the closeness with which they are packed in its crystal structure. Consider the iron oxide magnetite, with a density of 5.2 g/cm^3. This high density results partly from the high

Minerals and the Biological World

The crystal habit and chemical composition of some minerals have made them important in the biological world. The simplest examples are the calcium carbonate minerals calcite and aragonite, used as shell materials by a wide range of invertebrate animals such as clams and oysters. We have only to feel our own bones to recognize the importance of apatite, a calcium phosphate mineral that makes up vertebrate bone.

SUMMARY

What is a mineral? Minerals, the building blocks of rocks, are naturally occurring, inorganic solids with specific crystal structures and chemical compositions that either are fixed or vary within a defined range. A mineral is constructed of atoms, the small units of matter that combine in chemical reactions. An atom is composed of a nucleus of protons and neutrons, surrounded by electrons. The atomic number of an element is the number of protons in its nucleus, and its atomic mass is the sum of the masses of its protons and neutrons.

How do atoms combine to form the crystal structures of minerals? Chemical substances react with one another to form compounds either by gaining or losing electrons to become ions or by sharing electrons. The ions in a chemical compound are held together by ionic bonds, which form by electrostatic attraction between positive ions (cations) and negative ions (anions). Atoms that share electrons to form a compound are held together by covalent bonds. When a mineral crystallizes, atoms or ions come together in the proper proportions to form a crystal structure, which is an orderly three-dimensional geometric array in which the basic arrangement is repeated in all directions.

What are the major rock-forming minerals? Silicates, the most abundant minerals in Earth's crust, are crystal structures built of silicate tetrahedra linked in various ways. Tetrahedra may be isolated (olivines) or in single chains (pyroxenes), double chains (amphiboles), sheets (micas), or frameworks (feldspars). Carbonate minerals are made of carbonate ions bonded to calcium or magnesium or both. Oxide minerals are compounds of oxygen and metallic elements. Sulfide and sulfate minerals are composed of sulfur atoms in combination with metallic elements.

What are the physical properties of minerals? A mineral's physical properties, which indicate its composition and structure, include hardness—the ease with which its surface is scratched; cleavage—its ability to split or break along flat surfaces; fracture—the way in which it breaks along irregular surfaces; luster—the nature of its reflection of light; color—imparted by transmitted or reflected light to crystals, irregular masses, or a streak (the color of a fine powder); density—the mass per unit volume; and crystal habit—the shapes of individual crystals or aggregates.
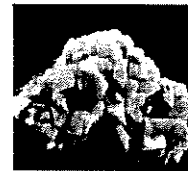
Minerals You Should Know

amphibole   calcite   diamond
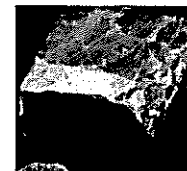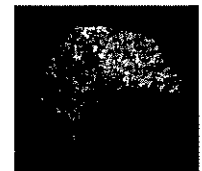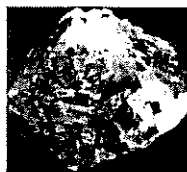
dolomite   feldspar   garnet
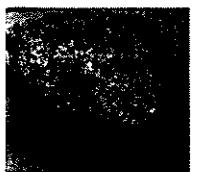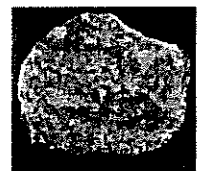
gypsum   halite   hematite

kaolinite   mica   olivine

pyrite   pyroxene   quartz

[Row 1: George Whiteley/Photo Researchers; Chip Clark; Charles O'Rear/Corbis. Row 2: Charles D. Winters/Photo Researchers; Chip Clark; Andrew Romeo–coolrox.com. Row 3: All by Chip Clark. Row 4: Vincent Cronin; Chip Clark; Chip Clark. Row 5: All by Chip Clark.]

Rocks: Records of Geologic Processes

"The silly question is the first intimation of some totally new development."

ALFRED NORTH WHITEHEAD

A rock is a naturally occurring solid aggregate of minerals. Some rocks, such as white marble, are composed of just one mineral, in this case calcite. A few rocks are composed of nonmineral matter. These include the noncrystalline, glassy volcanic rocks, obsidian and pumice, and coal, which is compacted plant remains. In an aggregate, minerals are joined so they retain their individual identity (**Figure 4.1**). What determines the physical appearance of a rock? They vary in color, in the sizes of their crystals or grains, and in the kinds of minerals that make them up. Along a road cut, for example, we might find a smooth black rock composed of volcanic glass and crystals of pyroxene and feldspar, particles that are too small to be seen with the naked eye (**Figure 4.2**). Nearby we might see a brownish rock with many large glittering crystals of mica and some grains of quartz and feldspar. Overlying both the black rock and the brown one we might see the remains of a former beach: horizontal layers of light-brown rock that appear to be made up of sand grains cemented together.

The appearance of a rock is determined partly by its mineralogy and partly by its texture. **Mineralogy**—the relative proportions of a rock's constituent minerals—helps determine how it looks, as well as other properties, as you will recall from Chapter 3. So does a rock's **texture**—the sizes and shapes of its mineral crystals and the way they are put together. These crystals (or grains), only a few millimeters in diameter in most rocks, are categorized as *coarse*, if they are large enough to be seen with the naked eye, and *fine*, if they are not. The mineralogy and texture that determine a rock's appearance are themselves determined by the rock's geologic origin—where and how it was formed (see Figure 4.2).

The dark rock in our road cut, called basalt, was formed by a volcanic eruption. Its mineralogy and texture depend on the chemical composition of rocks that were melted deep in the Earth. All rocks that form by the solidification of molten rock are called **igneous rocks.**
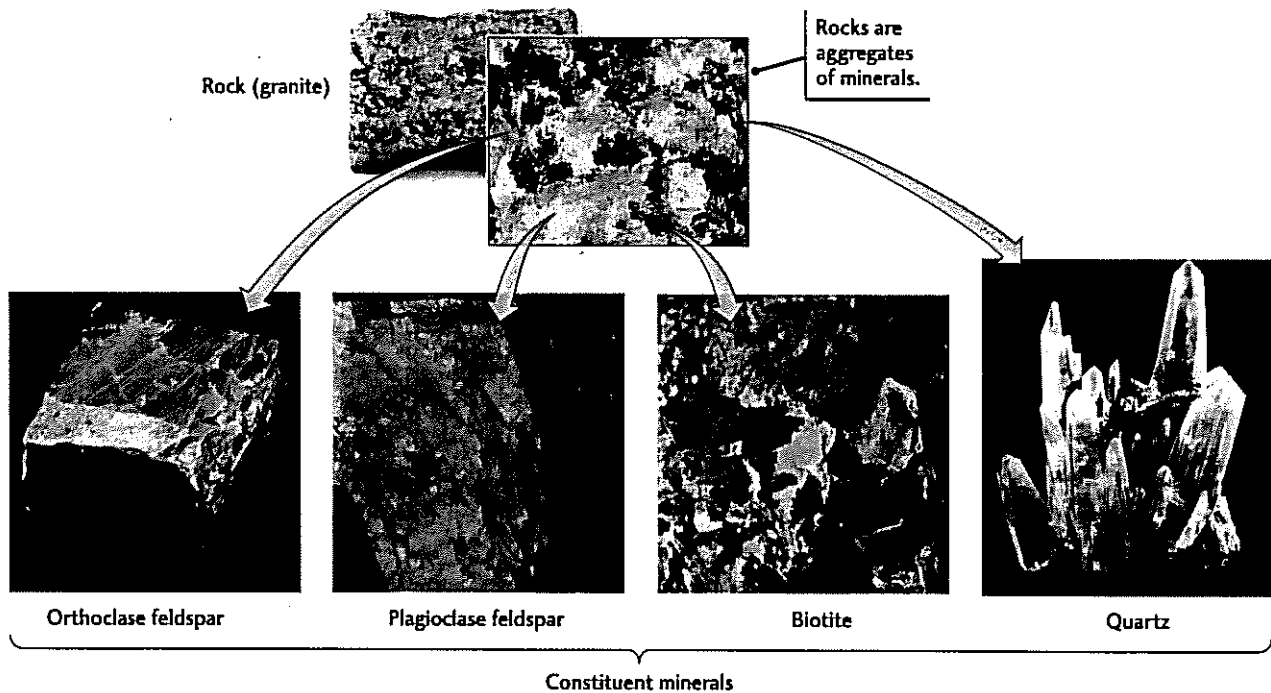
Figure 4.1 A rock is a naturally occurring aggregate of minerals. [Clockwise from top: J. Ramezani; J. Ramezani; José Manuel Sanchis Calvete/Corbis; Martin Miller/Visuals Unlimited; Arthur Hill/Visuals Unlimited; Chip Clark.]

The light-brown layered rock of the road cut, a sandstone, was formed as sand particles accumulated, perhaps on a beach, and eventually were covered over, buried, and cemented together. All rocks formed as the burial products of layers of sediments (such as sand, mud, and calcium carbonate shells), whether they were laid down on the land or under the sea, are called **sedimentary rocks.**

The brownish rock of our road cut, a schist, contains crystals of mica, quartz, and feldspar. It formed deep in Earth's crust as high temperatures and pressures transformed the mineralogy and texture of a buried sedimentary rock. All rocks formed by the transformation of preexisting solid rocks under the influence of high pressure and temperature are called **metamorphic rocks.**

A geologist's primary aim is to understand rock properties and to deduce their geologic origins from these properties. Such deductions further our understanding of the planet on which we live and provide important information about fuel reserves and solutions to environmental problems. For example, knowing that oil forms in certain kinds of sedimentary rocks that are rich in organic matter allows us to explore for new oil reserves more intelligently. Similarly, our knowledge of the properties of rocks will help us find new reserves of other useful and economically valuable mineral and energy resources, such as gas, coal, and metal ores.

Understanding how rocks form also guides us in solving environmental problems. Will this rock be prone to earthquake-triggered landslides? How might it transmit polluted waters in the ground? The underground storage of radioactive and other wastes depends on analysis of the rock to be used as a repository.

This chapter gives an overview of how geologists interpret the clues to understanding Earth provided by the three great families of rock: igneous, sedimentary, and metamorphic.

If rocks are the clues to many of the things we want to know about our planet, how do we go about interpreting them? We need a key, just as historians needed the Rosetta stone to crack the "code" of Egyptian hieroglyphs before they could read the inscriptions on temples and tombs. The first step in finding this key is to recognize the various kinds of rocks. The second step is to understand what their characteristics tell us about the surface and subsurface conditions under which they formed.

We will see what the appearance, texture, mineralogy, and chemical composition of a rock reveal about how and where it formed. We will look at how rock patterns found in subsurface drilling and in outcrops can help us reconstruct geologic history. Finally, we will trace the **rock cycle**—the set of geologic processes that convert each type of rock into
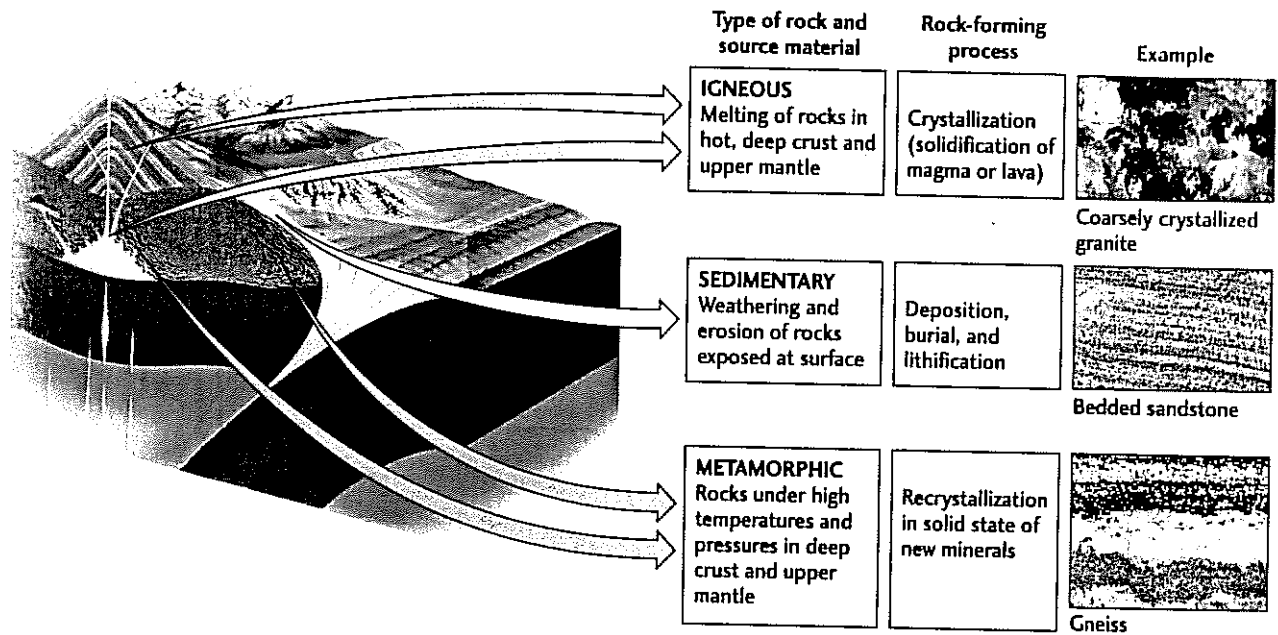
Figure 4.2 The minerals and textures of the three great rock groups are formed in different places in and on Earth by different geologic processes. As a result, geologists use mineralogical and chemical analyses to determine the origins of rocks and the processes that formed them. Granite, composed of quartz, feldspar, and mica crystals. [J. Ramezani.] Bedded sedimentary rock, made up of sandstones. [Breck P. Kent.] This crumpled and deformed metamorphic rock is a gneiss. [Breck P. Kent.]

the other two types—and see how these processes are all driven by plate tectonics and climate.

Igneous Rocks

Igneous rocks (from the Latin *ignis*, meaning "fire") form by crystallization from a magma, a mass of melted rock that originates deep in the crust or upper mantle. Here temperatures reach the 700°C or more needed to melt most rocks. When a magma cools slowly in the interior, microscopic crystals start to form. As the magma cools below the melting point, some of these crystals have time to grow to several millimeters or larger before the whole mass is crystallized as a coarse-grained igneous rock. But when a magma erupts from a volcano onto Earth's surface, it cools and solidifies so rapidly that individual crystals have no time to grow gradually. In that case, many tiny crystals form simultaneously, and the result is a fine-grained igneous rock. Geologists distinguish two major types of igneous rocks—intrusive and extrusive—on the basis of the sizes of their crystals.

Intrusive Igneous Rocks

Intrusive igneous rocks crystallize when magma intrudes into unmelted rock masses deep in Earth's crust. Large crystals grow as the magma cools, producing coarse-grained rocks. Intrusive igneous rocks can be recognized by their interlocking large crystals, which grew slowly as the magma gradually cooled (Figure 4.3). *Granite* is an intrusive igneous rock.

Extrusive Igneous Rocks

Extrusive igneous rocks form from rapidly cooled magmas that erupt at the surface through volcanoes. Extrusive igneous rocks, such as *basalt*, are easily recognized by their glassy or fine-grained texture (see Figure 4.3).

Common Minerals

Most of the minerals of igneous rocks are silicates, partly because silicon is so abundant and partly because many silicate minerals melt at the high temperatures and pressures
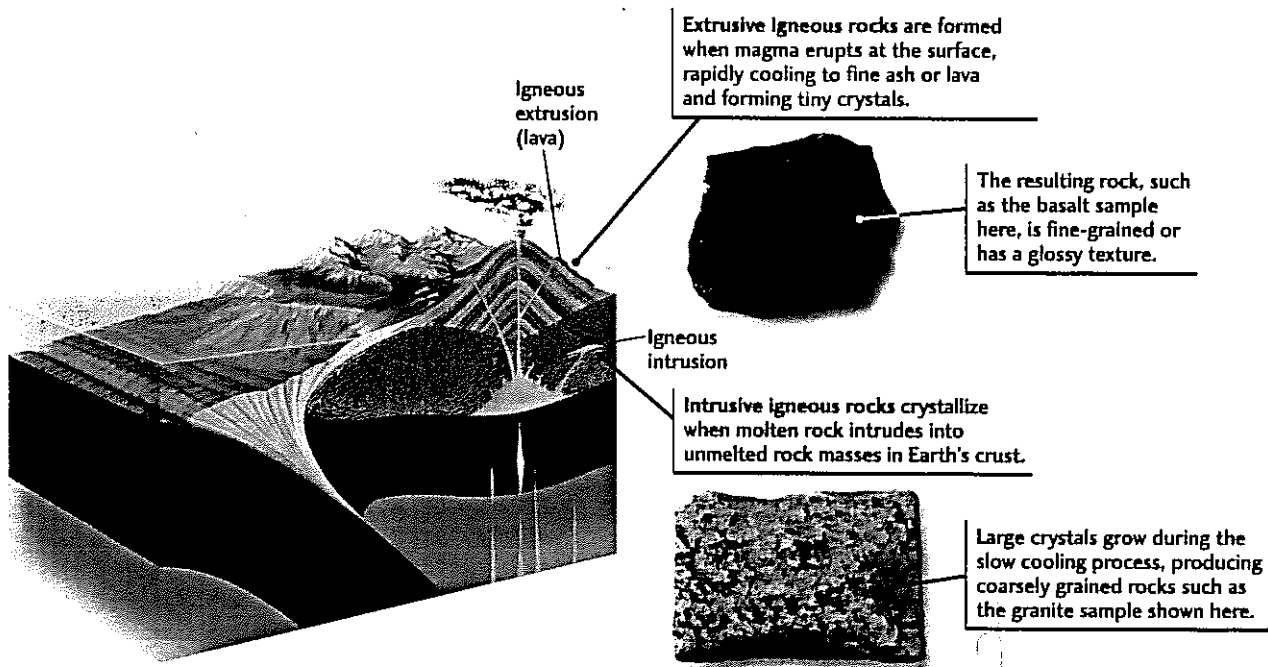
Igneous
extrusion
(lava)

Extrusive igneous rocks are formed
when magma erupts at the surface,
rapidly cooling to fine ash or lava
and forming tiny crystals.

The resulting rock, such
as the basalt sample
here, is fine-grained or
has a glossy texture.

Igneous
intrusion

Intrusive igneous rocks crystallize
when molten rock intrudes into
unmelted rock masses in Earth's crust.

Large crystals grow during the
slow cooling process, producing
coarsely grained rocks such as
the granite sample shown here.

Figure 4.3 The formation of *extrusive igneous rocks* (basalt is shown here) [Chip Clark] and *intrusive igneous rocks* (granite is shown here) [J. Ramezani].

reached in deeper parts of the crust and in the mantle. The common silicate minerals found in igneous rocks include quartz, feldspar, mica, pyroxene, amphibole, and olivine (Table 4.1).

Table 4.1 Some Common Minerals of Igneous, Sedimentary, and Metamorphic Rocks

Igneous Rocks	Sedimentary Rocks	Metamorphic Rocks
Quartz*	Quartz*	Quartz*
Feldspar*	Clay minerals*	Feldspar*
Mica*	Feldspar*	Mica*
Pyroxene*	Calcite	Garnet*
Amphibole*	Dolomite	Pyroxene*
Olivine*	Gypsum	Staurolite*
	Halite	Kyanite*

An asterisk indicates that a mineral is a silicate.

Sedimentary Rocks

Sediments, the precursors of sedimentary rocks, are found on Earth's surface as layers of loose particles, such as sand, silt, and the shells of organisms. These particles form at the surface as rocks undergo weathering and erosion. **Weathering** is all of the chemical and physical processes that break up and decay rocks into fragments of various sizes. The fragmented rock particles are then transported by **erosion,** the set of processes that loosen soil and rock and move them to the spot where they are deposited as layers of sediment (**Figure 4.4**). Weathering and erosion produce two types of sediments:

• **Clastic sediments** are physically deposited particles, such as grains of quartz and feldspar derived from a weathered granite. (*Clastic* is derived from the Greek word *klastos,* meaning "broken.") These sediments are laid down by running water, wind, and ice and form layers of sand, silt, and gravel.

• **Chemical and biochemical sediments** are new chemical substances that form by precipitation when some of a rock's components dissolve during weathering and are carried in river waters to the sea. These sediments include layers of such minerals as halite (sodium chloride) and calcite (calcium carbonate, most often found in the form of reefs and shells).

1 Particles of rock created by weathering...

2 ...are transported downhill by erosion...

3 ...and deposited as layers of sediment on land or in water,...

Beach

Continental margin

Igneous rock

Delta

Coral reefs

Continental shelf

4 ...where they form parallel layers, or bedding.

5 Buried sediments lithify by compaction and cementation.

6 Clastic sediments are made of deposited particles of sand, silt, and gravel.

7 Chemical and biochemical sediments are precipitated from seawater or made by coral reefs and shells.

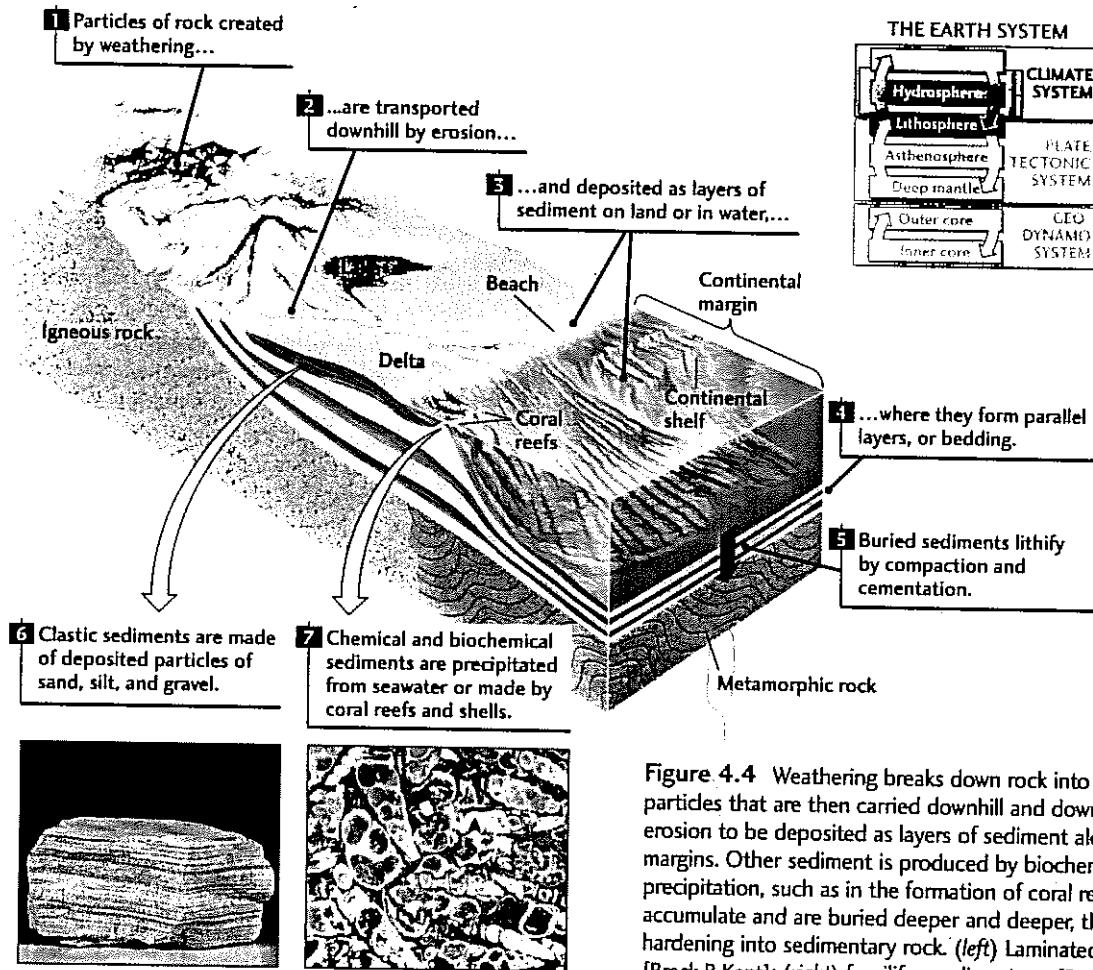Metamorphic rock

THE EARTH SYSTEM

Figure 4.4 Weathering breaks down rock into smaller particles that are then carried downhill and downstream by erosion to be deposited as layers of sediment along continental margins. Other sediment is produced by biochemical precipitation, such as in the formation of coral reefs. As layers accumulate and are buried deeper and deeper, they lithify, hardening into sedimentary rock. (left) Laminated sandstone [Breck P. Kent]; (right) fossiliferous limestone [Peter Kresan].

From Sediment to Solid Rock

Lithification is the process that converts sediments into solid rock, and it occurs in one of two ways:

• By compaction, as grains are squeezed together by the weight of overlying sediment into a mass denser than the original.

• By cementation, as minerals precipitate around deposited particles and bind them together.

Sediments are compacted and cemented after burial under additional layers of sediment. Thus sandstone forms by the lithification of sand particles, and limestone forms by the lithification of shells and other particles of calcium carbonate.

Layers of Sediment

Sediments and sedimentary rocks are characterized by bedding, the formation of parallel layers of sediment as particles settle to the bottom of the sea, a river, or a land surface.

Because sedimentary rocks are formed by surface processes, they cover much of Earth's land surface and seafloor. Although most rocks found at Earth's surface are sedimentary, their volume is small compared to the igneous and metamorphic rocks that make up the main volume of the crust because they are difficult to preserve (Figure 4.5).

Common Minerals

The common minerals of clastic sediments are silicates, because silicate minerals predominate in rocks that weather to form sedimentary particles (see Table 4.1). The most abundant minerals in clastic sedimentary rocks are quartz, feldspar, and clay minerals.

The most abundant minerals of chemically or biochemically precipitated sediments are carbonates, such as calcite, the main constituent of limestone. Dolomite, also found in limestone, is a calcium-magnesium carbonate formed by precipitation during lithification. Two other chemical sediments—gypsum and halite—form by precipitation as seawater evaporates.
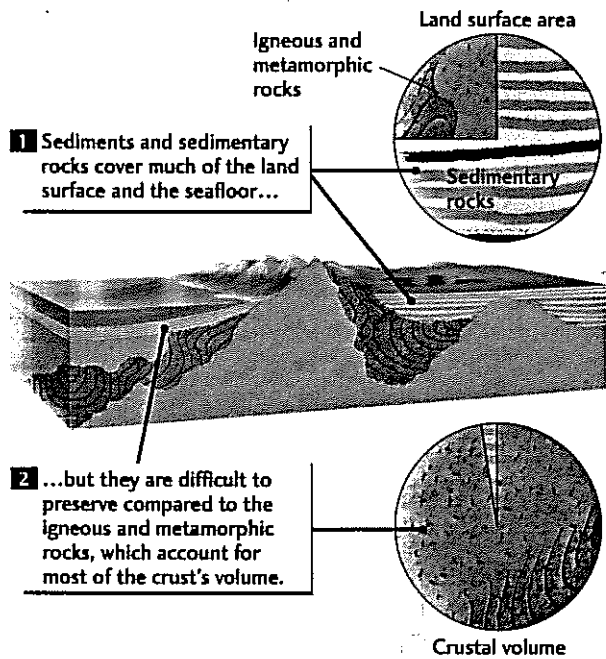
Figure 4.5 Sediments and sedimentary rocks cover much of Earth's land surface and the seafloor.

Metamorphic Rocks

Metamorphic rocks take their name from the Greek words for "change" (*meta*) and "form" (*morphe*). These rocks are produced when high temperatures and pressures deep in the Earth cause any kind of rock—igneous, sedimentary, or other metamorphic rock—to change its mineralogy, texture, or chemical composition while maintaining its solid form. The temperatures of metamorphism are below the melting points of the rocks (about 700°C) but high enough (above 250°C) for the rocks to change by recrystallization and chemical reactions.

Regional and Contact Metamorphism

Metamorphism may take place over a widespread area or a limited one (Figure 4.6). Regional metamorphism occurs where high pressures and temperatures extend over large regions, as happens where plates collide. Regional metamorphism accompanies plate collisions that result in mountain building and the folding and breaking of sedimentary layers that were once horizontal. Where high temperatures are restricted to smaller areas, such as the rocks near and in contact with an intrusion, rocks are transformed by **contact metamorphism.**

Many regionally metamorphosed rocks, such as schists, have characteristic **foliation,** wavy or flat planes produced when the rock was structurally deformed into folds. Granu-

lar textures are more typical of most contact metamorphic rocks and of some regional metamorphic rocks formed by very high pressure and temperature.

Common Minerals

Silicates are the most abundant minerals of metamorphic rocks because the parent rocks are also rich in silicates (see Table 4.1). Typical minerals of metamorphic rocks are quartz, feldspar, mica, pyroxene, and amphibole—the same kinds of silicates characteristic of igneous rocks. Several other silicates—kyanite, staurolite, and some varieties of garnet—are characteristic of metamorphic rocks alone. These minerals form under conditions of high pressure and temperature in the crust and are not characteristic of igneous rocks. They are therefore good indicators of metamorphism. Calcite is the main mineral of marbles, which are metamorphosed limestones.

Pressure-Temperature-Time Paths

Plate tectonics makes regional metamorphism a dynamic process in which volumes of rock are subjected to changing conditions of pressure and temperature over time. Consequently, regionally metamorphosed rocks contain distinctive mineral assemblages in which earlier assemblages are overprinted by later assemblages. Such rocks thus record regimes of pressure and temperature which change in time. **Pressure-temperature-time paths** are recorded not only by changes in the assemblages of minerals but also by changes in the chemical compositions of the minerals themselves. Metamorphic pressure-temperature paths are discussed in detail in Chapter 9.

Where We See Rocks

Rocks are not found in nature conveniently divided into separate bodies—igneous here, sedimentary there, metamorphic in another place. Instead, we find them arranged in patterns determined by the geologic history of a region. Geologists map these patterns both at the surface and as projected into the interior, and they try to deduce the geologic past from the present variety and distribution of the rocks.

If we were to drill a hole into any spot on Earth, we would find rocks that represent the geologic history of that region. In the top few kilometers of most regions, we would probably find sedimentary rock. Drilling deeper, perhaps 6 to 10 km down, we would eventually penetrate an underlying area of older igneous and metamorphic rock.

In fact, thousands of relatively shallow holes have been drilled on the continents in the search for oil, water, and mineral resources. These holes are major sources of information, mainly about sedimentary rocks and their history. In the quest for more data on the deep continental crust, the governments of several countries—including the United States,

Volcanism

*"What a book a devil's chaplain might write on the clumsy,
wasteful, blundering, low, and horribly cruel works of nature."*

CHARLES DARWIN

I magine a volcanic eruption in which an area of ground the size of New York City collapses, a region larger than Vermont is buried under hot ash that snuffs out all life, and fields 1000 or 2000 km away are blanketed by 20 cm of ash and rendered infertile. Imagine that the volcanic dust thrown high into the stratosphere dims the Sun for a year or two, so that there are no summers. Unbelievable? Yet it has happened at least twice in what is now the United States: at Yellowstone in Wyoming 600,000 years ago and at Long Valley, California, 760,000 years ago. This was long before humans first reached North America 30,000 years ago, but not very long ago on the 4.5-billion-year geologic time scale. We know that these events took place because volcanic rocks formed by the eruptions have been identified and dated.

A large portion of Earth's crust, oceanic and continental, is made up of volcanic rock. Beginning as magma deep inside the Earth, volcanic rock serves as a kind of window through which we can dimly perceive Earth's interior.

In this chapter, we examine volcanism, the process by which magma from the interior of the Earth rises through the crust, emerges onto the surface as lava, and cools into hard volcanic rock. We will look at the major types of lava, eruptive styles and the landforms they create, and the kinds of environmental disruption that volcanoes can cause. We will see how plate tectonics and mantle convection can explain why the majority of volcanoes occur at plate boundaries and a few occur at "hot spots" within plates. We will give examples of how volcanoes interact with other components of the Earth system, particularly the atmosphere, the oceans, and the biosphere. Finally, we will consider ways to mitigate the destructive potential of volcanoes and benefit from their chemical riches and heat energy.

Ancient philosophers were awed by volcanoes and their fearsome eruptions of molten rock. In their efforts to explain volcanoes, they spun myths about a hot, hellish underworld below Earth's surface. Basically, they had the right idea. Modern researchers, using science rather than mythology, also see in volcanoes evidence of Earth's internal heat.

lavas and the rocks that they form vary according to the magmas from which they were derived. In Chapter 5, we saw that igneous rocks and their precursor magmas can be divided into three major groups—felsic, intermediate, and mafic—on the basis of their chemical composition (see Table 5.2). Igneous rocks are further classified as intrusive (they cooled slowly below the surface and are coarse-grained as a result) or extrusive (they cooled quickly on the surface and are fine-grained). The major intrusive igneous rocks are granite (felsic), diorite (intermediate), and gabbro (mafic). The major extrusive counterparts are rhyolite (felsic), andesite (intermediate), and—most common of all—basalt (mafic). These classifications are summarized in Figure 5.4. With this framework in mind, let us examine the types of lava and how they flow and solidify.

Types of Lava

The several types of lava leave behind different landforms: volcanic mountains that vary in shape and solidified lava flows that vary in character. These variations result from differences in the chemical composition, gas content, and temperature of lavas. The higher the silica content and the lower the temperature, for example, the more viscous (more resistant to flow) the lava is and the more slowly it moves. The more gas a lava contains, the more violent its eruption is likely to be.

Basaltic Lavas Basaltic lava, dark in color, erupts at 1000° to 1200°C—close to the temperature of the upper mantle. Because of its high temperature and low silica content, basaltic lava is extremely fluid and can flow downhill fast and far. Streams flowing as fast as 100 km/hour have been observed, although velocities of a few kilometers per hour are more common. In 1938, two daring Russian volcanologists measured temperatures and collected gas samples while floating down a river of molten basalt on a raft of colder solidified lava. The surface temperature of the raft was 300°C, and the river temperature was 870°C. Lava streams that traveled more than 50 km from their source have been witnessed in historical times.

Basaltic lava flows vary according to the conditions under which they erupt. Important examples include

• *Flood Basalts* Highly fluid basaltic lava that erupts on flat terrain can spread out in thin sheets as a flood of lava. Successive flows often pile up into immense basaltic lava plateaus, called **flood basalts,** such as those at the great Columbia Plateau of Oregon and Washington (**Figure 6.2**).

• *Pahoehoe and Aa* Cooling basaltic lava flowing downhill falls into one of two categories, according to its surface form: pahoehoe (pronounced pa-ho´-ee-ho´-ee) or aa (ah-ah). **Figure 6.3** shows examples of both.

> *Pahoehoe* (the word is Hawaiian for "ropy") forms when a highly fluid lava spreads in sheets and a thin, glassy elastic skin congeals on its surface as it cools. As
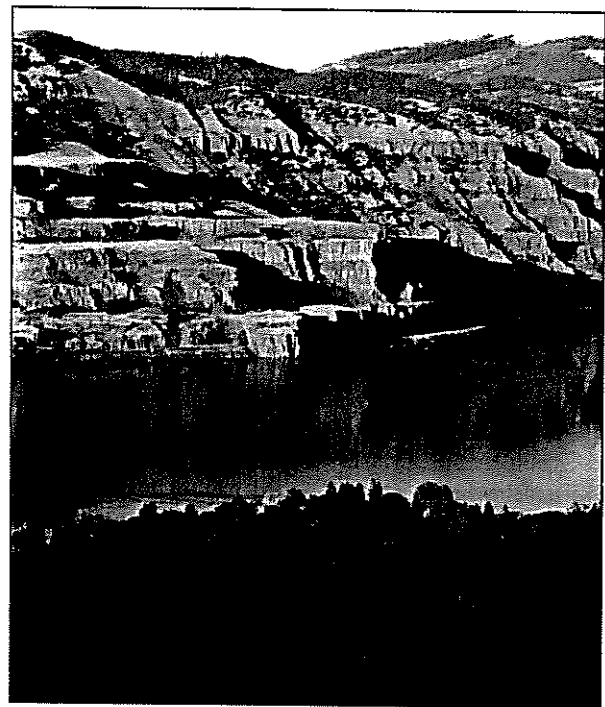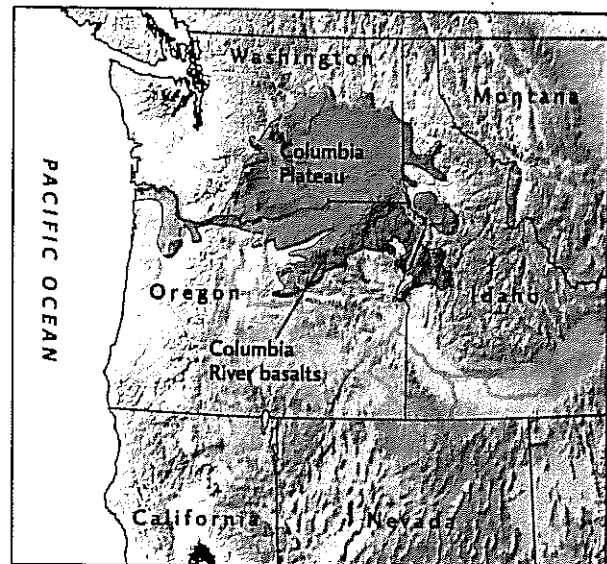
Figure 6.2 View of Columbia Plateau, Washington. Successive flows of flood basalts piled up to build this immense plateau, which covers a large area of Washington and Oregon. [Dave Schiefelbein.]

the molten liquid continues to flow below the surface, the skin is dragged and twisted into coiled folds that resemble rope.

"*Aa*" is what the unwary exclaim after venturing barefoot onto lava that looks like clumps of moist,

IV

Aa lava

Pahoehoe lava

`|_____|`
`~1 m`
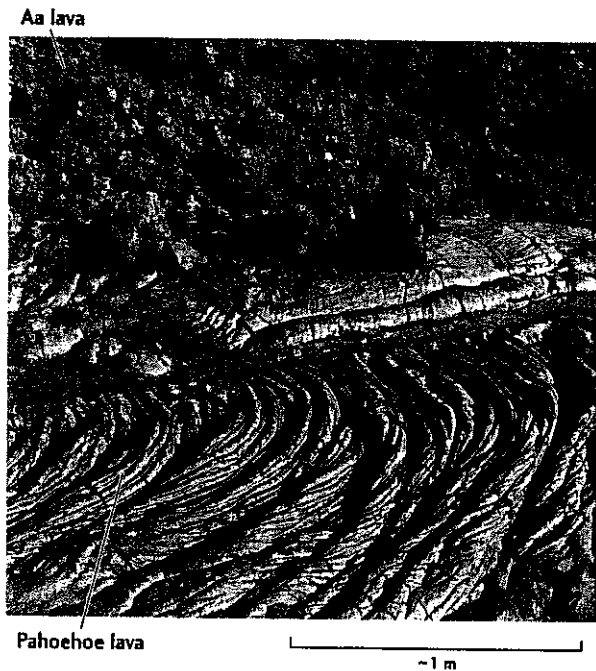
Figure 6.3 Two types of lava, ropy pahoehoe (*bottom*) and jagged blocks of aa (*top*). Mauna Loa volcano, Hawaii. [Kim Heacox/DRK.]

freshly plowed earth. Aa is lava that has lost its gases and consequently is more viscous than pahoehoe. Thus, it moves more slowly, which allows a thick skin to form. As the flow continues to move, the thick skin breaks into rough, jagged blocks. The blocks ride on the viscous, massive interior, piling up a steep front of angular boulders that advances like a tractor tread. Aa is truly treacherous to cross. A good pair of boots may last about a week on it, and the traveler or geologist can count on cut knees and elbows.

A single downhill basaltic flow commonly has the features of pahoehoe near its source, where the lava is still fluid and hot, and of aa farther downstream, where the flow's surface—having been exposed to cool air longer—has developed a thicker outer layer.

• *Pillow Lavas* A geologist who comes across **pillow lavas**—piles of ellipsoidal, pillowlike blocks of basalt about a meter wide—knows that they formed in an underwater eruption (see Figure 5.13) even if they are now on dry land. In fact, pillow lavas are an important indicator that a region was once under water. Geologist-divers have actually observed pillow lavas forming on the ocean floor off Hawaii. Tongues of molten basaltic lava develop a tough, plastic skin on contact with the cold ocean water. Because lava inside the skin cools more slowly, the pillow's interior develops a crystalline texture, whereas the quickly chilled skin solidifies to a crystal-less glass.

Rhyolitic Lavas Rhyolite, the most felsic lava, is light in color. It has a lower melting point than basalt and erupts at temperatures of 800° to 1000°C. It is much more viscous than basalt because of its lower temperature and higher silica content. Rhyolite moves 10 or more times more slowly than basalt, and, because it resists flow, it tends to pile up in thick, bulbous deposits.
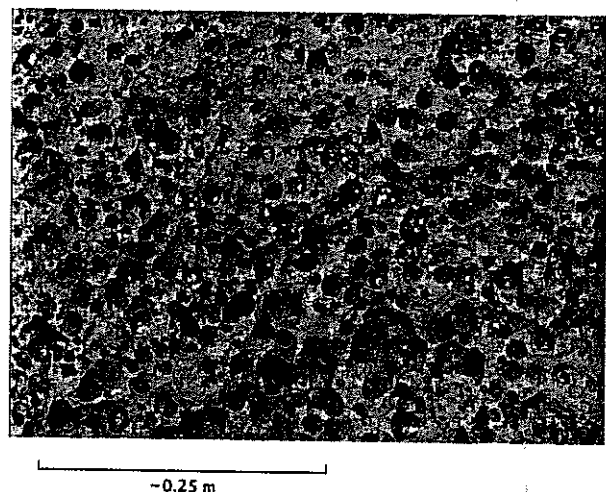
Andesitic Lavas Andesite, with an intermediate silica content, has properties that fall between those of basalts and rhyolites.
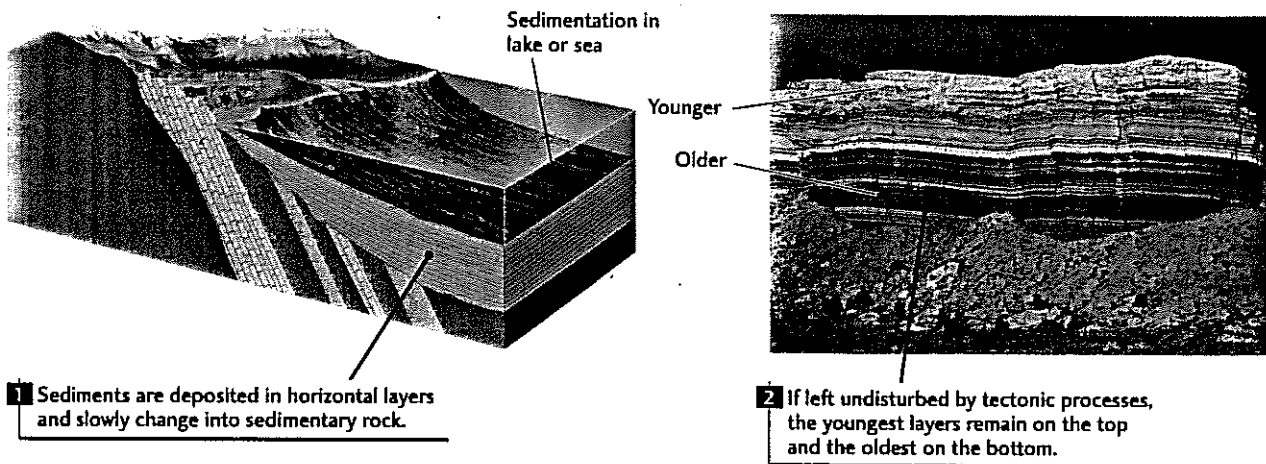
Textures of Lavas

Lavas have other features that reflect the temperatures and pressures under which they formed. They can have a glassy or fine-grained texture, such as that of obsidian, if they cool quickly. Coarse-grained textures, such as those of volcanic tuffs, result if they cool slowly beneath the surface. They can also contain little bubbles, created when pressure falls suddenly as the lava rises and cools. Lava is typically charged with gas, like soda in an unopened bottle. When lava rises, the pressure on it decreases, just as the pressure on a soda drops when its bottle cap is removed. And just as the carbon dioxide in the soda creates bubbles when it is released, the water vapor and other dissolved gases escaping from lava create gas cavities, or *vesicles* (Figure 6.4). A frothy texture in solidified lava provides geologists with details of the rock's volcanic origins. An extremely vesicular volcanic rock, usually rhyolitic in composition, is called *pumice*. Some pumice has so much void space that it is light enough to float on water.

Pyroclastic Deposits

Water and gases in magmas can have even more dramatic effects on eruptive styles. Before a magma erupts, the con-

`|_____|`
`~0.25 m`

Figure 6.4 Vesicular basalt sample. [Glenn Oliver/Visuals Unlimited.]

Sedimentation in lake or sea

Younger

Older

1 Sediments are deposited in horizontal layers and slowly change into sedimentary rock.

2 If left undisturbed by tectonic processes, the youngest layers remain on the top and the oldest on the bottom.

Figure 10.3 Layers of sedimentary rock in Marble Canyon, part of the Grand Canyon. The Grand Canyon was cut by the Colorado River through what is now northern Arizona. These layers record millions of years of geologic history. [Fletcher & Baylis/Photo Researchers.]

stratigraphic succession is the chronology of the constituent beds and the sedimentary conditions that they imply.

With a geologic timekeeper, or "stratigraphic clock," geologists can tell whether one rock layer is older than another, although they cannot necessarily tell how much older in years. One might expect that a stratigraphic succession would provide a measure of time in actual years if sediments had accumulated continuously at a steady rate, had compacted a constant amount as they lithified, and had not eroded. If we knew that muddy sediments accumulated at a rate of 10 m per million years, for instance, then 100 m of mudstone would represent 10 million years of deposition.

In practice, however, we cannot accurately gauge time in years from stratigraphy, for several reasons. First, as we saw in Chapter 8, sediments do not accumulate at a constant rate in any sedimentary environment. During a flood, a river may deposit a layer of sand several meters thick in its channel over just a few days, whereas in the years between floods, it will deposit a sand layer only a few centimeters thick. Even on the deep ocean floor, where it may take 1000 years for a layer of mud 1 mm thick to accumulate, sedimentation is unsteady and the thickness of sediment cannot be used for precise timekeeping. In addition, the rate at which sediment is deposited varies widely in different sedimentary environments.

The second reason that stratigraphy is an imprecise timekeeper is that the rock record does not tell us how many years have passed between periods of deposition. Many places on the floor of a river valley receive sediment only during times of flood. The times between floods are not represented by any sediment. Over the course of Earth's history, in various places there have been long intervals, some lasting *millions* of years, in which no sediments were de-

posited at all. In other places and at other times, sedimentary rocks may have been removed by erosion. Although we often can tell where a gap in the record exists, we rarely can say exactly how long an interval it represents.

The final reason—and the one most important to geologists who wish to compare the geologic histories of different parts of the Earth—is that stratigraphy alone cannot be used to determine the relative ages of two widely separated beds. A geologist might be able to establish the relative age of one bed or a series of beds by following an outcrop a limited distance, but there is no way of knowing whether a rock layer in Arizona, say, is older or younger than one in northern Canada.

Those early geologists found that fossils were the key to detecting missing time intervals and correlating the relative ages of rocks at different geographic locations. Fossils became the single most important tool for constructing an accurate geologic time scale for the entire planet.

Fossils as Timepieces

To a great many students today, it must be obvious that fossils are the remains of ancient organisms. Some look very much like animals now living, although others—such as the trilobites in the photograph at the beginning of this chapter—are the remains of extinct life-forms. Fossils can be shells, teeth, bones, impressions of plants, or tracks of animals (**Figure 10.4**). The most common fossils in rocks of the past half-billion years are the shells of invertebrate animals, such as clams, oysters, and the ammonite group shown in Figure 10.4a. Much less common are the bones of vertebrates, such as mammals, reptiles, and dinosaurs. Yet fossil dinosaur bones do exist, and they have told us much

(a)

(b)

Figure 10.4 Animal and plant fossils. (a) Ammonite fossils, ancient examples of a large group of invertebrate organisms that are now largely extinct. Their sole representative in the modern world is the chambered nautilus. [Chip Clark.] (b) Petrified Forest, Arizona. These ancient logs are millions of years old. Their substance was completely replaced by silica, which preserved all the original details of form. [Tom Bean.]

about the nature of these long-extinct beasts. Plant fossils are abundant in some rocks, particularly those associated with coal beds, where leaves, twigs, branches, and even whole tree trunks can be recognized (Figure 10.4b). Fossils are not found in intrusive igneous rocks, because the original biological material is lost in a hot melt. Fossils are rarely found in metamorphic rocks, because any remains of organisms are usually too changed and distorted to be recognized.

The ancient Greeks were probably the first to surmise that fossils are records of ancient life, but it was not until modern times that the concept took hold and its consequences were explored. One of the first modern thinkers to make the connection between fossils and once-living organisms was Leonardo da Vinci, in the fifteenth century. In the seventeenth century, Nicolaus Steno compared what he called "tongue-stones" found in the Mediterranean region with similarly shaped teeth of modern sharks and concluded that the stones were the remains of ancient life.

By the end of the eighteenth century, after hundreds of fossils and their relationships to modern organisms had been described and catalogued, the evidence that fossils are the remains of formerly living creatures was overwhelming. Thus **paleontology,** the study of the history of ancient life from the fossil record, took its place beside geology, the study of Earth's history from the rock record.

The dividends from the study of fossils did not accrue to geology alone. Young Charles Darwin's famous voyage as naturalist on the *Beagle* (1831–1836) vastly extended his knowledge of the great variety of fossil organisms and of what their presence in rocks portended. He also had an opportunity to see a host of unfamiliar animal and plant species in their native habitats. In 1859, Darwin proposed the theory of evolution. It revolutionized scientific thinking about the origins of the millions of species of animal and plant life and provided a sound theoretical framework for paleontology.

Well before Darwin, in 1793, William Smith, a surveyor working in southern England, had recognized that fossils

0 Second
Rupture expands circularly on fault plane, sending out seismic waves in all directions.


Focus
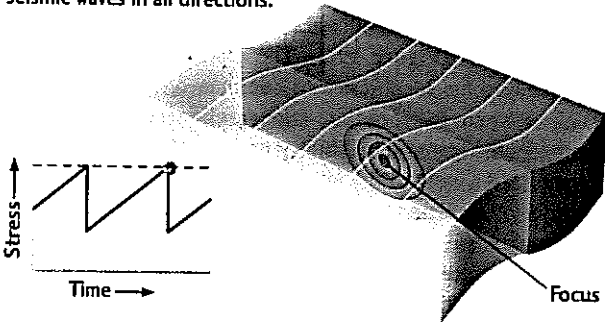
5 Seconds
Rupture continues to expand as a crack along the fault plane. When rupture front reaches the surface, displacements occur along the fault trace and rocks at the surface begin to rebound from their deformed state.

Fault cracks at surface
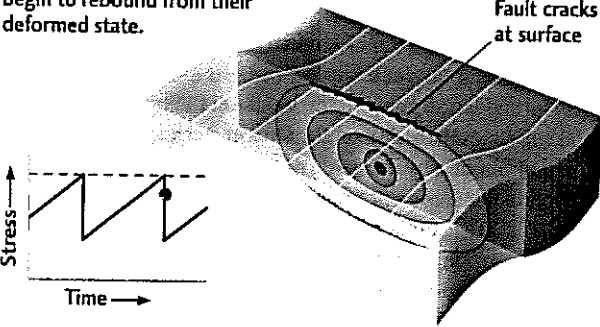
10 Seconds
Rupture front progresses down the fault plane, reducing the stress and allowing rocks on either side to rebound. Seismic waves continue to be emitted in all directions as the fault propagates.

Fault crack extends
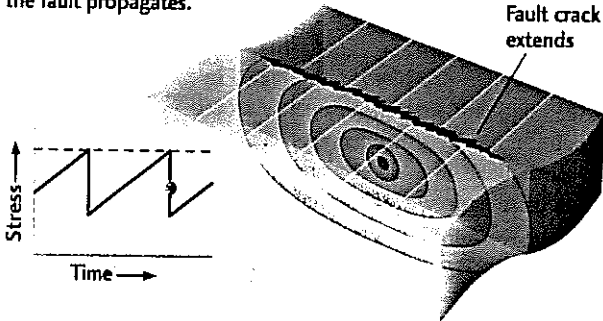
20 Seconds
Rupture has progressed along the entire length of the fault. The fault has reached its maximum displacement, and the earthquake stops.
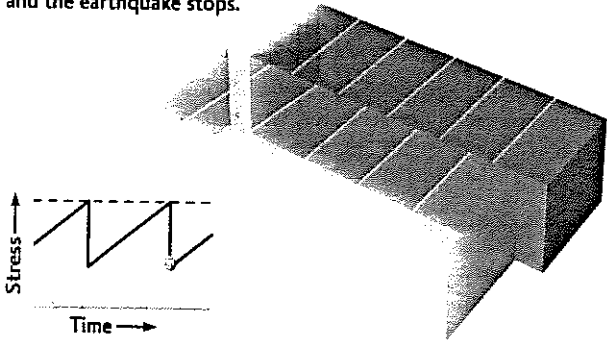

more complex than the simplest models of plate tectonics might suggest. These complexities contribute to the difficulty of predicting earthquakes. We will find that earthquakes cannot yet be predicted reliably, although their destructiveness can be reduced considerably. To do the latter, we must use our geological knowledge of where large earthquakes are likely to occur in designing buildings, dams, bridges, and other structures to withstand earthquake shaking.

What Is an Earthquake?

We have seen that plate movements generate forces in narrow zones at the boundaries between tectonic plates. These global forces leave their imprint locally in ways that can be described by the concepts of stress, strain, and strength. *Stress* is the local force per unit area that causes rocks to deform. *Strain* is a measure of the amount of deformation. Rocks fail—that is, they lose cohesion and break into two or more parts—when they are stressed beyond a critical value called the *strength.* Brittle rock formations commonly fail on faults when they are stressed beyond their strength. An **earthquake** occurs when rocks being stressed suddenly break along a new or preexisting fault. The two blocks of rock on either side of the fault slip suddenly, setting off ground vibrations, or **seismic waves** (from the Greek *seismos,* meaning "shock" or "earthquake"), that are often destructive. When the fault slips, the stress is reduced, dropping to a level below the rock strength. After the earthquake, the stress begins to increase again, and the cycle is repeated (Figure Story 19.1).

Figure 19.2 Fault rupture during an earthquake begins at the focus and spreads out over the fault plane, radiating seismic waves wherever the fault is slipping. The sequence shown here occurs in a few tens of seconds between Time 3 and Time 4 in Figure Story 19.1. The photo shows a 3-m right-lateral offset of the fence, near Bolinas, California, following the 1906 San Francisco earthquake. Displacement of the fence was observed to be similar to the fence at Time 4 in elastic rebound sequence. [G. K. Gilbert]
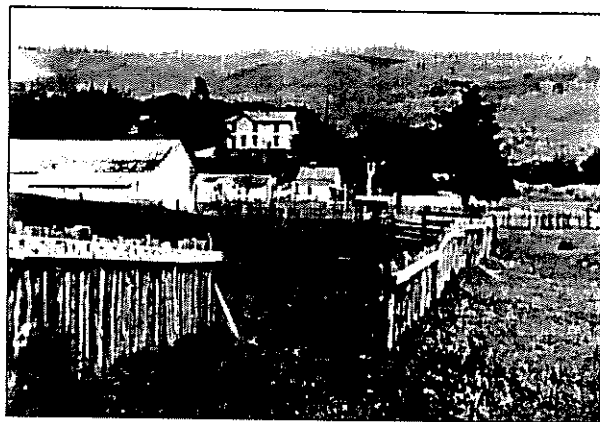
Earthquakes occur most often at plate boundaries, where stresses are concentrated and strain is intense.

Elastic Rebound Explains Why Earthquakes Occur

The earthquake on the San Andreas fault that devastated San Francisco in 1906 received the most detailed study of any earthquake up to that time. A geologist who investigated that catastrophe, Henry Fielding Reid of Johns Hopkins University, advanced the **elastic rebound theory** to explain why earthquakes occur.

To visualize what happens in an earthquake, imagine the following experiment carried out across a fault between two hypothetical crustal blocks. Suppose that surveyors had painted straight lines on the ground, running perpendicular to a fault and extending from one block to the other, as in Figure Story 19.1. The two blocks are being pushed in opposite directions by plate motions. The weight of the overlying rock presses them together, however, so friction locks them together along the fault. They do not move, just as a car does not move when the emergency brake is engaged. Instead of slipping along the fault as stress builds up, the blocks are strained elastically near the fault, as shown by the bent lines in Figure Story 19.1. By *elastically,* we mean that the blocks would spring back and return to their undeformed, stress-free shape if the fault were suddenly to unlock.

As the slow plate movements continue to push the blocks in opposite directions, the strain in the rocks—evidenced by the bending of the survey lines—continues to build up, for decades or even centuries. At some point, the strength of the rocks is exceeded. The frictional bond that locks the fault can no longer hold somewhere along the fault, and it breaks. The blocks slip suddenly, and the rupture extends over a section of the fault. Figure Story 19.1 shows how the two blocks have rebounded—sprung back to their undeformed state—after the earthquake. The imaginary bent lines have straightened, and the two blocks have been displaced. The distance of the displacement is called the **slip.**

Fault Rupture During Earthquakes

The point at which the slip begins is the **focus** of the earthquake (see Figure Story 19.1). The **epicenter** is the geographic point on Earth's surface directly above the focus. For example, you might hear in a news report: "Seismologists at the California Institute of Technology report that the epicenter of last night's destructive earthquake in California was located 56 kilometers southeast of Los Angeles. The depth of the focus was 10 kilometers."

For most earthquakes occurring in continental crust, the focal depths range from about 2 to 20 km. Continental earthquakes are rare below 20 km, because under the high temperatures and pressures found at greater depths, the crust deforms as a ductile material and cannot support brittle fracture (just as hot wax flows when stressed, whereas cold wax breaks; see Chapter 11). In subduction zones, however,

where cold oceanic lithosphere plunges back into the mantle, earthquakes can begin at depths as great as 690 km.

The fault rupture does not occur all at once. It begins at the focus and expands outward on the fault plane, typically at 2 to 3 km/s (Figure 19.2). The rupture stops where the stresses become insufficient to continue breaking the fault (such as where the rocks are stronger) or where the rupture enters ductile material in which it can no longer propagate as a fracture. As we will see later in this chapter, the size of an earthquake is related to the total area of fault rupture. Most earthquakes are very small, with rupture dimensions much less than the depth of focus, and so the rupture never breaks the surface. In large, destructive earthquakes, however, surface breaks are common. For example, the great 1906 San Francisco earthquake caused surface displacements averaging about 4 m along a 400-km section of the San Andreas fault (see Figure 19.2). Faulting in the largest earthquakes can extend as far as 1000 km, and the slip of the two blocks can be as large as 10 m. Generally, the longer the fault rupture, the larger the fault slip.

The sudden slipping of the blocks at the time of the earthquake reduces the stress on the fault and releases much of the stored strain energy. Most of this stored energy is converted to frictional heating of the fault zone, but part of it is released as seismic waves that travel outward from the rupture, much as waves ripple outward from the spot where a stone is dropped into a still pond. The focus of the earthquake generates the first seismic waves, although slipping parts of the fault continue to generate waves until the rupture stops. In a large event like the 1906 San Francisco earthquake, the propagating fault continues to produce waves for many tens of seconds. These waves can cause damage all along the fault break, even far from the epicenter.
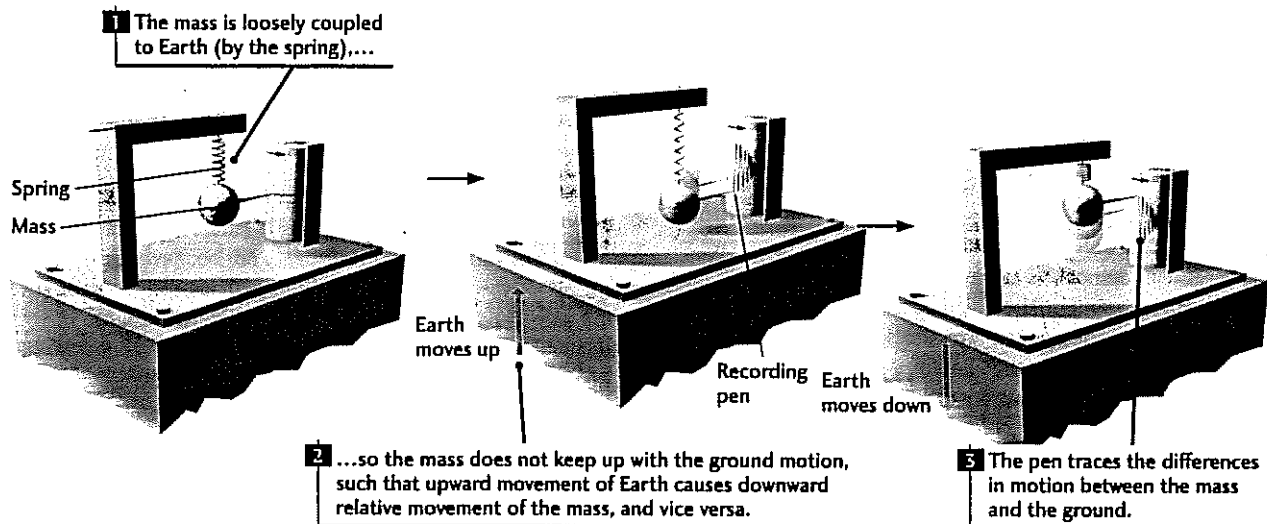
The elastic strain energy that slowly builds up over decades when two blocks are pushed in opposite directions is analogous to the strain energy stored in a rubber band when it is slowly stretched. The sudden release of energy in an earthquake, signaled by slip along a fault and the generation of seismic waves, is analogous to the violent rebound or springback that occurs when the rubber band breaks. The elastic energy stored in the stretched rubber band is suddenly released in the backlash. In the same way, elastic energy accumulates and is stored for many decades in rocks under stress. The energy is released at the moment the fault ruptures, and some of it is radiated as seismic waves in the few minutes of an earthquake.

In its idealized form, the elastic rebound model implies that faults should exhibit a periodic buildup and release of strain energy. Sequences of real earthquakes rarely exhibit this simple behavior, however, which is one of the reasons earthquakes are so difficult to predict.
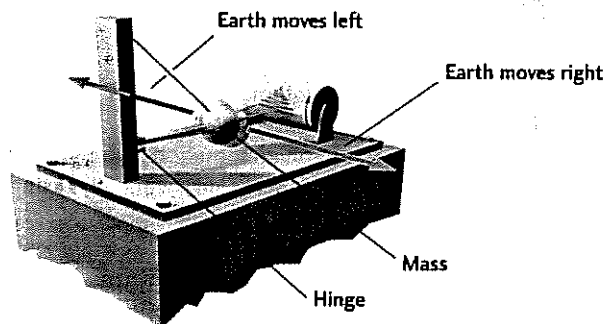
Foreshocks and Aftershocks

An example of the complexities not described by simple elastic rebound is the phenomenon of aftershocks. An after-

(a) Seismograph designed to detect vertical movement

1 The mass is loosely coupled to Earth (by the spring),...

Spring

Mass

Earth moves up

Recording pen

Earth moves down

2 ...so the mass does not keep up with the ground motion, such that upward movement of Earth causes downward relative movement of the mass, and vice versa.

3 The pen traces the differences in motion between the mass and the ground.

(b) Seismograph designed to detect horizontal movement

Earth moves left

Earth moves right

Mass

Hinge

Figure 19.4 Seismographs record (a) vertical or (b) horizontal motion. Because of its loose coupling to the Earth through the spring (a) or hinge (b) and its inertia, the mass does not keep up with the motion of the ground. The pen traces the differences in motion between the mass and the ground, in this way recording the vibrations of seismic waves. A typical observatory has instruments to measure three components of ground motion: up-down, horizontal east-west, and horizontal north-south.

through Earth's interior. Finally come the **surface waves,** which travel around Earth's surface (Figure Story 19.5).

P waves in rock are analogous to sound waves in air, except that P waves travel through the solid rock of Earth's crust at about 6 km/s, which is about 20 times faster than sound waves travel through air. Like sound waves, P waves are *compressional waves,* so called because they travel through solid, liquid, or gaseous materials as a succession of compressions and expansions. P waves can be thought of as push-pull waves: they push or pull particles of matter in the direction of their path of travel.

S waves travel through solid rock at a little more than half the speed of P waves. They are called *shear waves* because they displace material at right angles to their path of travel. Shear waves do not exist in liquids or gases.

Surface waves are confined to Earth's surface and outer layers because, like waves on the ocean, they need a free surface to ripple. Their speed is slightly less than that of S waves. One type of surface wave sets up a rolling motion in the ground; another type shakes the ground sideways (see Figure Story 19.5).

People have felt seismic waves and witnessed their destructiveness throughout history, but not until the close of the nineteenth century were seismologists able to devise instruments to record them. Seismic waves enable us to locate earthquakes and determine the nature of faulting, and they provide our most important tool for probing Earth's deep interior.

Locating the Epicenter

Locating a quake's epicenter is analogous to deducing the distance to a lightning bolt on the basis of the time interval between the flash of light and the sound of thunder—the greater the distance to the bolt, the larger the time interval. Light travels faster than sound, so the lightning flash may be likened to the P waves of earthquakes and the thunder to the slower S waves.

The time interval between the arrival of P and S waves depends on the distance the waves have traveled from the focus. This relationship is established by recording seismic waves from an earthquake or underground nuclear explosion

about the consequences for our habitat and for the generations to come.

In our increasingly systematic search of the Earth for new sources of the fuel and minerals on which we depend, we use our geological knowledge of how known natural deposits form to determine where we may find more of them. At the same time, we are becoming more sensitive to the finiteness of Earth's resources and the delicacy of its environment. We are paying more attention to extracting and using natural resources more efficiently, without damaging the environment. We are beginning to think about how we can change our use of resources to achieve sustainable development—development that will preserve the prospects of future generations.

Resources and Reserves

Two major questions arise in all discussions of materials that we draw from the Earth: How much is left? How long will it last? The amount left consists of more (we hope) than reserves. Reserves are deposits that have already been discovered and that can be mined economically and legally at the present time. Resources, in contrast, constitute the entire amount of a given material that may become available for use in the future. Resources include reserves, plus discovered but currently unprofitable deposits, plus undiscovered deposits that geologists think they may be able to find eventually (Figure 22.1). Often, resources that are too poor in quality or quantity to be worth mining now or that are too difficult to retrieve become profitable when new technology is developed or prices rise. A recent example is the discovery and production of oil and gas from large reservoirs in the continental margin of the Gulf of Mexico, in water depths reaching 3000 m.
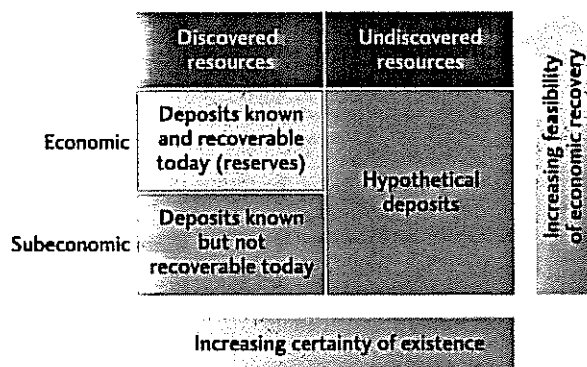
Figure 22.1 Categories that constitute total resources. Discovered resources consist of reserves—known deposits that are economically minable today—and deposits that are known but that are currently subeconomic. Undiscovered resources are hypothetical deposits that we may find.

Reserves are considered a dependable measure of supply as long as economic and technological conditions remain as they are now. As conditions change, some resources become reserves, and vice versa. The conversion of oil resources in the North Sea into productive oil fields, for example, was accelerated by new technology and by price increases resulting from political instability in the oil-exporting nations of the Middle East. The assessment of resources is much less certain than the assessment of reserves. Any figure cited as the resources of a particular material represents only an educated guess of how much will be available in the future.

Most useful geological materials are considered nonrenewable because geological processes produce them more slowly than civilization uses them up. Coal and oil, for example, will be exhausted faster than nature can replenish them. This fact makes it increasingly important to develop renewable resources such as solar energy, which is essentially infinite in supply, and fuels such as ethanol, which is derived from crops that can be replanted after they are harvested.

Energy Resources

Energy is fundamental to everything. A crisis in the supply of energy can bring a modern society to a halt. Wars have been fought over access to supplies of fuel resources; economic recession and destructive currency inflation have resulted from gyrations in the price of oil. It is not surprising that energy is the biggest business in the world.

Fuel resources are measured in units appropriate to the material, for example, barrels of oil, tons of coal. To make it easier to compare the energy available from different fuel resources, a common unit called the quad is used. A quad is a measure of the energy that can be extracted from a given amount of fuel. The quad is based on a standard measure of energy called the British thermal unit (Btu). One Btu is the amount of energy needed to raise the temperature of 1 pound of water by 1°F. One quad equals 10^{15} Btu. The United States uses about 98 quads of energy a year. Figure 22.2 shows one estimate of the world's remaining nonrenewable energy resources of all types—about 360,000 quads.

In 1999, the world consumption of energy was 382 quads, of which all but about 35 quads were nonrenewable. (Consumption is estimated to grow to 607 quads by 2020.) Calculations based on these numbers can be deceptive, however. For example, simply dividing total resources by annual consumption might lead us to conclude (mistakenly) that many hundreds of years of resources remain before we have to worry about depletion of the supply. However, some energy sources will give out before others, and the various sources of energy are not readily interchangeable. Also, as we will see later in this chapter and in Chapter 23, each of

in chemical terms, is the combustion of organic matter consisting of compounds of carbon and hydrogen. Organic matter is produced by plants and animals. The organic matter in this case is a tree, which obtained its energy for growth by a process called photosynthesis (see Chapter 23). During photosynthesis, plants use the energy supplied by sunlight to convert carbon dioxide and water to carbohydrates. Thus, we can look upon a piece of wood or any piece of plant matter as a photosynthetic product that can be returned by burning or decay to the carbon dioxide and water from which it was made.

If we burn wood that was buried and transformed 300 million years ago into the combustible rock known as coal, we are using energy stored by photosynthesis from late Paleozoic sunlight. We are recovering "fossilized" energy. Crude oil and natural gas were also created by a process of burial and chemical transformation of dead organic matter into a combustible liquid and gas, respectively. We refer to all such resources derived from natural organic materials, from coal to oil and natural gas, as **fossil fuels** (Figure 22.4).
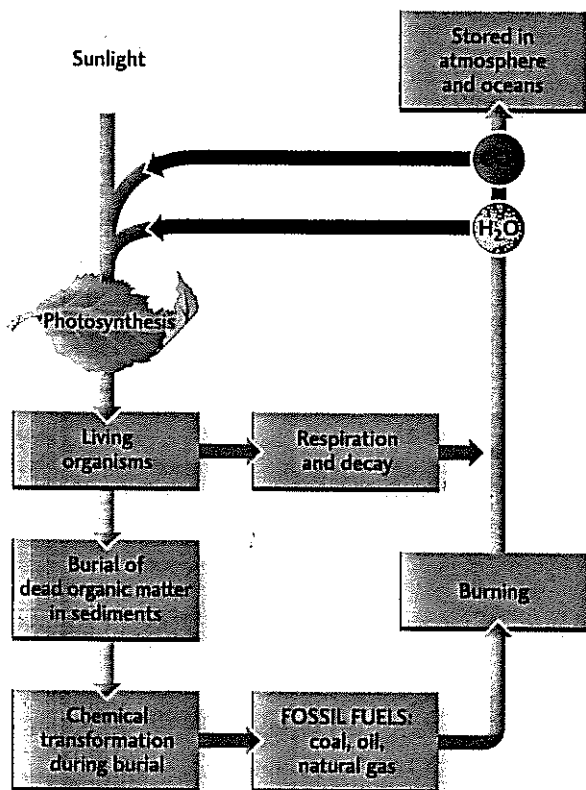

Figure 22.4 Photosynthesis produces organic matter from carbon dioxide (CO_2) and water (H_2O). If dead organic matter is buried and transformed into coal, oil, or natural gas, it becomes a fossilized product of photosynthesis—a fossil fuel. The burning of fossil fuels releases the carbon dioxide and water from which they were made.

More than 85 percent of the world's energy is presently derived from fossil fuels.

Oil and Natural Gas

Crude oil (petroleum) and natural gas in minable form develop under special environmental and geological conditions. Both are the organic debris of former life—plants, bacteria, algae and other microorganisms that have been buried, transformed, and preserved in marine sediments.

How Oil and Gas Form

Oil and gas begin to form when more organic matter is produced than is destroyed by scavengers and decay. This condition exists in environments where the production of organic matter is high—such as in the coastal waters of the sea, where large numbers of organisms thrive—and the supply of oxygen in bottom sediments is inadequate to decompose all the organic matter. Many offshore sedimentary basins on continental shelves satisfy both these conditions. In such environments, and to a lesser degree in some river deltas and inland seas, the rate of sedimentation is high and organic matter is buried and protected from decomposition.

During millions of years of burial, chemical reactions triggered by the elevated temperatures at depth slowly transform some of the organic material into liquid and gaseous compounds of hydrogen and carbon (hydrocarbons). The hydrocarbons are the combustible materials of oil and natural gas. Compaction of muddy organic sediments in their source beds forces the hydrocarbon-containing fluids and gases into adjacent beds of permeable rock (such as sandstones or porous limestones), which we call *oil reservoirs*. The low density of oil and gas causes them to rise to the highest place that they can reach, where they float atop the water that almost always occupies the pores of permeable formations.

Oil Traps Geological conditions that favor the large-scale accumulation of oil and natural gas are combinations of structure and rock types that create an impermeable barrier to upward migration—an **oil trap.** Some oil traps are caused by a structural deformation and are called *structural traps.* One type of structural trap is formed by an anticline in which an impermeable shale overlies a permeable bed of sandstone (Figure 22.5a). The oil and gas accumulate at the crest of the anticline—the gas highest, the oil next—both floating on the groundwater that saturates the sandstone. Similarly, an angular unconformity or displacement at a fault may place a dipping permeable limestone bed opposite an impermeable shale, creating a structural trap for oil (Figure 22.5b). Other oil traps are created by the original pattern of sedimentation, such as when a dipping permeable sandstone bed thins out against an impermeable shale (Figure 22.5c). These are called *stratigraphic traps.* Oil can

Weathering and Erosion

*"The heights of our land are thus leveled with the shores;
our fertile plains are formed from the ruins of mountains."*

JAMES HUTTON (1788)

Solid as the hardest rocks may seem, all rocks—like rusting old automobiles and yellowed old newspapers— eventually weaken and crumble when exposed to water and the gases of the atmosphere. Unlike those cars and newspapers, however, rocks may take thousands of years to disintegrate. In this chapter, we look closely at two geologic processes that break down and fragment rocks: weathering and erosion.

Weathering is the general process by which rocks are broken down at Earth's surface. Weathering produces all the clays of the world, all soils, and the dissolved substances carried by rivers to the ocean. Rocks weather in two ways:

• **Chemical weathering** occurs when the minerals in a rock are chemically altered or dissolved. The blurring or disappearance of lettering on old gravestones and monuments is caused mainly by chemical weathering.

• **Physical weathering** takes place when solid rock is fragmented by mechanical processes that do not change its chemical composition. The rubble of broken stone blocks and columns that were once stately temples in ancient Greece is primarily the result of physical weathering. Physical weathering also caused the cracks and breaks in the ancient tombs and monuments of Egypt.

Chemical and physical weathering reinforce each other. Chemical decay weakens fragments of rocks and makes them more susceptible to breakage. The smaller the pieces produced by physical weathering, the greater the surface area available for chemical weathering.

Weathering, Erosion, and the Rock Cycle

After tectonics and volcanism have made mountains, chemical decay and physical breakup join with rainfall, wind, ice, and snow to wear away those mountains. **Erosion** is the set of processes that loosen and transport soil and rock downhill or

downwind. These processes carry away weathered material on Earth's surface and deposit it elsewhere. As erosion moves weathered solid material, it exposes fresh, unaltered rock to weathering.

Weathering and erosion are major geologic processes in the rock cycle and in Earth systems, as described in Chapter 4. Along with tectonics and volcanism (two other elements of the rock cycle), weathering and erosion shape Earth's surface and alter rock materials, converting igneous and other rocks into sediment and forming soil. In some instances, weathering and erosion are inseparable. When a rock such as pure limestone or rock salt is weathered by rainwater, for example, all the material is completely dissolved and carried away in the water as ions in solution. Chemically weathered material contributes most of the dissolved matter in the oceans.

The early sections of this chapter emphasize chemical weathering, because it is in some ways the fundamental driving force of the whole process. The effects of physical weathering, always important, depend largely on chemical decay. First, however, we examine the factors that control weathering.

Why Do Some Rocks Weather More Rapidly Than Others?

All rocks weather, but the manner and rate of their weathering vary. The four key factors that control the fragmentation and decay of rocks are the properties of the parent rock, the climate, the presence or absence of soil, and the length of time the rocks are exposed to the atmosphere. These four factors are summarized in Table 7.1.

The Properties of the Parent Rock

The nature of a parent rock affects weathering because (1) various minerals weather at different rates and (2) a rock's structure affects its susceptibility to cracking and fragmentation. Old inscriptions on gravestones offer clear evidence of the varying rates at which rocks weather. The carved letters on a recently erected gravestone stand out in sharp relief from the stone's polished surface. After a hundred years in a moderately rainy climate, however, the surface of a limestone will be dull and the letters inscribed on it will have almost melted away, much as the name on a bar of soap disappears after a few washes (Figure 7.1). Granite, on the other hand, will show only minor changes. The differences in the weathering of granite and limestone result from their different mineral compositions. Given enough time, however, even a resistant rock will ultimately decay. After several hundred years, the granite monument will also have weathered appreciably, and its surface and letters will be somewhat dulled and blurred. If we used a hand lens to look more closely at the weathered granite, we would see different patterns of weathering in its constituent mineral grains. The feldspar crystals would show signs of corrosion, and their surfaces would be chalky and covered with a thin layer of soft clay. The chemical composition of the outer layers of the grains of feldspar would have changed, creating a new mineral. The quartz crystals would appear fresh—clear and unaltered.

Table 7.1 Major Factors Controlling Rates of Weathering

| | Weathering Rate | | |
Slow			Fast
PROPERTIES OF PARENT ROCK			
Mineral solubility in water	Low (e.g., quartz)	Moderate (e.g., pyroxene, feldspar)	High (e.g., calcite)
Rock structure	Massive	Some zones of weakness	Very fractured or thinly bedded
CLIMATE			
Rainfall	Low	Moderate	High
Temperature	Cold	Temperate	Hot
PRESENCE OR ABSENCE OF SOIL AND VEGETATION			
Thickness of soil layer	None—bare rock	Thin to moderate	Thick
Organic content	Low	Moderate	High
LENGTH OF EXPOSURE			
Short	Moderate	Long	

Figure 7.1 Early-nineteenth-century gravestones at Wellfleet, Massachusetts. The stone on the right is limestone and is so weathered that it is unreadable. The stone on the left is slate, which retains its legibility under the same conditions. [Courtesy of Raymond Siever.]

A rock's structure also affects physical weathering. Granite monuments may remain unbroken and uncracked even after centuries of exposure, though they may show evidence of some chemical weathering. Intrusive igneous rocks, including many granites, may be massive—that is, large masses that show no changes in rock type or structure. Massive rocks have no planes of weakness that contribute to cracking or fragmentation. In contrast, shale, a sedimentary rock that splits easily along thin bedding planes, breaks into small pieces so quickly that only a few years after a new road is cut through a shale, the rock will become rubble.

Climate: Rainfall and Temperature

A tour of graveyards across the North American continent, from the southern United States to northern Canada and Alaska, would reveal that the rate of both chemical and physical weathering varies not only with the properties of the rock but also with the climate—the amount of rainfall and the temperature. High temperatures and heavy rainfall increase the growth rate of organisms and thus promote chemical weathering. Cold and dryness impede the process. Old gravestones in hot, humid Florida are badly chemically weathered, but those of the same age in the equally hot but arid Southwest are hardly affected. And gravestones in cold, dry arctic regions show even less chemical weathering than those found in the Southwest. In cold climates, water can't dissolve minerals because it is frozen. In arid regions, water is relatively unavailable. In both cases, populations of or-

ganisms are at a minimum and chemical weathering proceeds slowly.

On the other hand, climates that minimize chemical weathering may enhance physical weathering. For example, freezing water may act as a wedge, widening cracks and pushing a rock apart.

The Presence or Absence of Soil

Soil, one of our most valuable natural resources, is composed of fragments of bedrock, clay minerals formed by the chemical alteration of bedrock minerals, and organic matter produced by organisms that live in the soil. Although soil is itself a product of weathering, its presence or absence affects the chemical and physical weathering of other materials. An old nail that has been buried in soil usually will be so badly rusted that you can snap it like a matchstick. Yet a nail pried from the wood of a centuries-old house may still be strong, covered with only a thin layer of rust. Similarly, a mineral in the soil of a lowland valley may be badly altered and corroded, whereas the same mineral exposed in a nearby cliff of bedrock will be much less weathered. Although the cliff is exposed to occasional rain, the bare rock is usually dry, and weathering proceeds very slowly. No soil forms on the cliff because rain quickly carries loosened particles down to lower areas, where they can accumulate.

Soil production is a *positive-feedback process*—that is, the product of the process advances the process itself. Once soil starts to form, it works as a geological agent to weather

rock more rapidly. The soil retains rainwater, and it hosts a variety of vegetation, bacteria, and other organisms. These life-forms create an acidic environment that, in combination with moisture, promotes chemical weathering, which alters or dissolves minerals. Plant roots and organisms tunneling through the soil promote physical weathering by helping to create fractures in a rock. Chemical and physical weathering, in turn, lead to the production of more soil.

The Length of Exposure

The longer a rock weathers, the greater its chemical alteration, dissolution, and physical breakup. Rocks that have been exposed at Earth's surface for many thousands of years form a rind—an external layer of weathered material ranging from several millimeters to several centimeters thick—that surrounds the fresh, unaltered rock. In dry climates, some rinds have grown as slowly as 0.006 mm per 1000 years.

Recently erupted volcanic lavas and ash deposits have had a very short period of exposure at Earth's surface and so are relatively unweathered. Because we know the dates of modern eruptions, we can measure the times required for various degrees of weathering to occur. In the years since the eruptions of Mount St. Helens in 1980, for example, the volcanic ash deposits have weathered appreciably and have altered to form other minerals. After the same length of time, masses of solidified lava are still relatively fresh. The difference in the extent of weathering is due mainly to the fact that the ash is made up of very small particles, which weather faster than the more massive volcanic rocks.

We now consider the two types of weathering—chemical and physical—in more detail.

Chemical Weathering

Rocks chemically weather when their constituent minerals react with air and water. In these chemical reactions, some minerals dissolve. Others combine with water and such components of the atmosphere as oxygen and carbon dioxide to form new minerals. We can deduce some chemical reactions from observations in the field. We can get a better picture of the mechanisms of chemical weathering if we combine field observations with laboratory experiments that simulate natural processes. We begin our investigation by examining the chemical weathering of feldspar, the most abundant mineral in Earth's crust.

The Role of Water in Weathering: Feldspar and Other Silicates

Feldspar is a key mineral in a great many igneous, sedimentary, and metamorphic rocks. Many other kinds of rock-forming silicate minerals also weather much as feldspar

does. Feldspar is one of many silicates that are altered by chemical reactions to form the water-containing minerals known as clay minerals. Feldspar's behavior during weathering helps us understand the weathering process in general, for two reasons:

1. There is an overwhelming abundance of silicate minerals in the Earth.

2. The chemical processes of dissolution and alteration that characterize feldspar weathering also characterize weathering in other kinds of minerals.

Earlier in this chapter, when we described the minerals one would find in a weathered granite gravestone, we noted that the feldspar crystals would be corroded and altered. A more extreme example of feldspar weathering can be found in granite boulders in soils of the humid tropics. There, many of the factors that promote weathering—heavy rainfall, high temperature, the presence of soil, and abundant organic activity—are present. Granite boulders found in the tropics are so weakened that they can be easily kicked or pounded into a heap of loose mineral grains. Most of the feldspar particles in these boulders have been altered to clay. Greatly magnified under an electron microscope, any remaining feldspar grains would display corrosion and be coated with a clay rind (Figure 7.2). Quartz crystals, in contrast, would be relatively intact and unaltered.
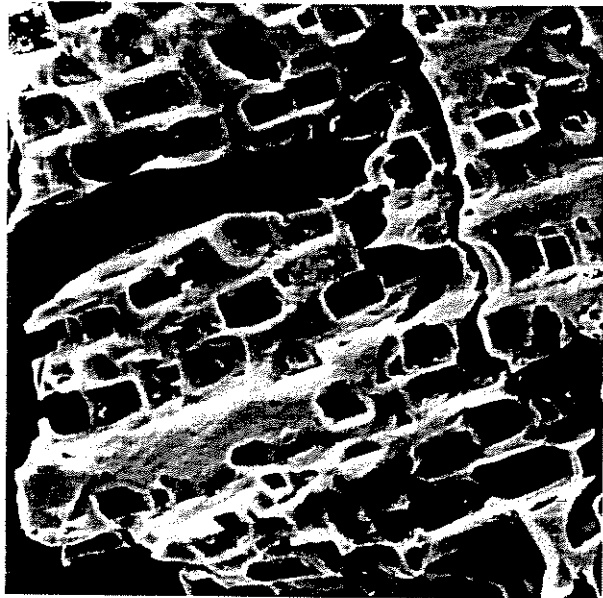

Figure 7.2 A scanning electron micrograph of feldspar etched and corroded by chemical weathering in soil. [From R. A. Berner and G. R. Holden, Jr., "Mechanism of Feldspar Weathering: Some Observational Evidence," Geology 5 (1977): 369.]
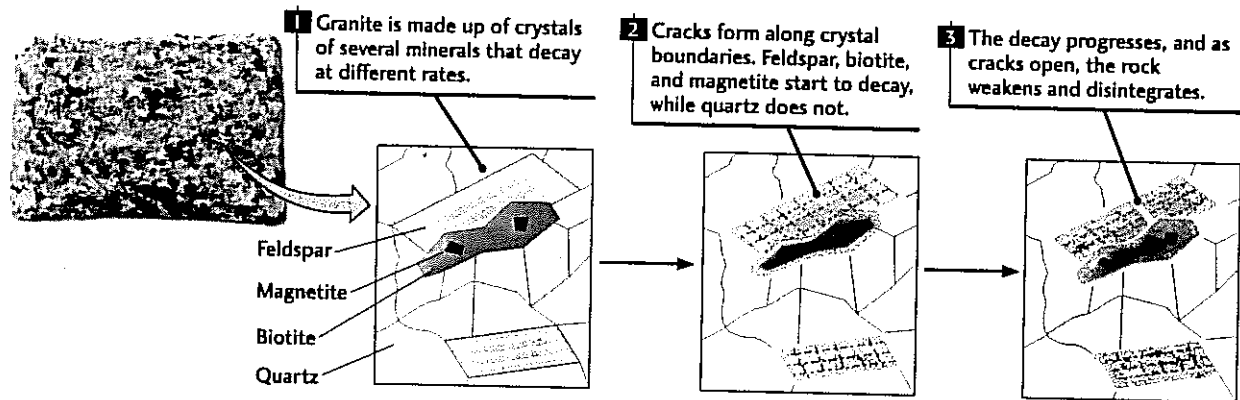
Figure 7.3 Diagrammatic microscopic views of stages in the disintegration of granite.
[Chip Clark.]

In a sample of unweathered granite, the rock is hard and solid because the interlocking network of quartz, feldspar, and other crystals holds it tightly together. When the feldspar is altered to a loosely adhering clay, however, the network is weakened and the mineral grains are separated (Figure 7.3; see also Figure 5.1). In this instance, chemical weathering, by producing the clay, also promotes physical weathering because the rock now fragments easily along widening cracks at mineral boundaries.

The white to cream-colored clay produced by the weathering of feldspar is **kaolinite,** named for Gaoling, a hill in southwestern China where it was first obtained. Chinese artisans had used pure kaolinite as the raw material of pottery and china for centuries before Europeans borrowed the idea in the eighteenth century.

Only in the severely arid climates of some deserts and polar regions does feldspar remain relatively unweathered. This observation points to water as an essential component of the chemical reaction by which feldspar becomes kaolinite. Kaolinite is a hydrous (that is, containing water in the crystal structure) aluminum silicate. In the reaction that produces kaolinite, the solid feldspar undergoes **hydrolysis** (*hydro* means "water" and *lysis* means "to loosen"). The feldspar is broken down and also loses several chemical components. The alteration is analogous to the chemical reaction that takes place when we make coffee. Solid coffee reacts chemically with hot water to make a solution—the liquid coffee. The reaction extracts caffeine and other components of the solid bean, leaving behind spent coffee grounds. Similarly, rainwater filters into the ground, altering feldspar in rock particles and leaving behind the kaolinite as a residue (**Figure 7.4**).

The only part of a solid that reacts with a fluid is the solid's surface, so as we increase the surface area of the solid, we speed up the reaction. For example, as we grind coffee beans into finer and finer particles, we increase the ratio of their surface area to their volume. The finer the cof-
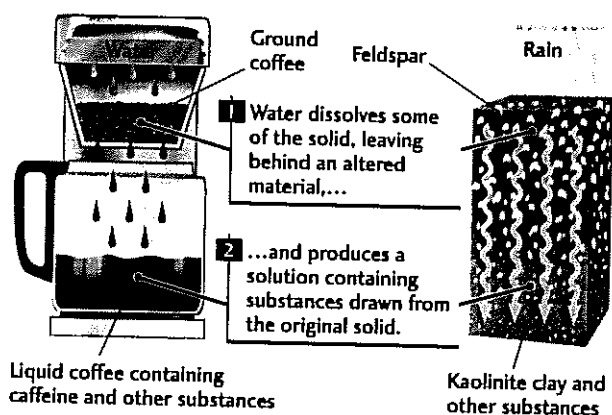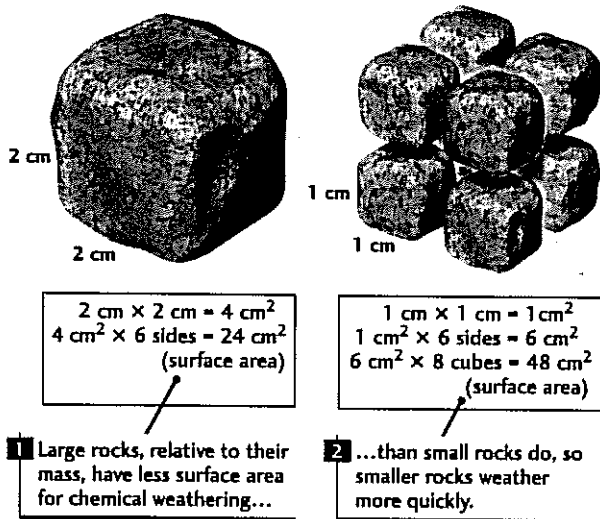

Figure 7.4 The process by which feldspar decays is analogous to the brewing of coffee.

fee beans are ground, the faster their reaction with water, and the stronger the brew becomes. Similarly, the smaller the fragments of minerals and rocks, the greater the surface area. The ratio of surface area to volume increases greatly as the average particle size decreases, as shown in **Figure 7.5**.

Dissolving Feldspar in Pure Water To learn more about the weathering of feldspar, we can perform a simple laboratory experiment: immerse feldspar in pure water and analyze the solution for the kinds of material that have dissolved. First we grind the common feldspar of granite, orthoclase ($KAISi_3O_8$), to a powder. This speeds the reaction by exposing more surface area to the water. When we analyze samples taken from the solution after some time has passed, we find small amounts of potassium and silica (SiO_2) dissolved in the water. The reaction of feldspar with water releases dissolved silica and dissolved potassium ions (K^+) and leaves behind

| 2 cm × 2 cm = 4 cm² |
| 4 cm² × 6 sides = 24 cm² |
| (surface area) |

| 1 cm × 1 cm = 1 cm² |
| 1 cm² × 6 sides = 6 cm² |
| 6 cm² × 8 cubes = 48 cm² |
| (surface area) |

1 Large rocks, relative to their mass, have less surface area for chemical weathering...

2 ...than small rocks do, so smaller rocks weather more quickly.

Figure 7.5 As a rock mass breaks into smaller pieces, more surface becomes available for the chemical reactions of weathering.

the new mineral kaolinite, $Al_2Si_2O_5(OH)_4$. We can now describe a chemical reaction for the weathering of granite by saying that feldspar reacts with water to form kaolinite.

Two major points about this reaction give us information about the gains and losses of material as feldspar weathers:

1. Potassium and silica are "lost" by the feldspar and appear as dissolved materials in the water solution.

2. Water is absorbed into the kaolinite crystal structure. This absorption of water is called *hydration* and is one of the major processes of weathering.

Carbon Dioxide, Weathering, and the Climate Geosystem

Variability in the atmosphere's concentration of carbon dioxide leads to corresponding variability in the rate of weathering (**Figure Story 7.6**). Higher levels of carbon dioxide in the atmosphere lead to higher levels in the soil, which increases the rate of weathering (Figure Story 7.6a). In addition, carbon dioxide, a greenhouse gas, makes Earth's climate warmer and thus promotes weathering. Weathering, in turn, converts carbon dioxide into bicarbonate ions (Figure Story 7.6b) and so decreases the amount of carbon dioxide in the atmosphere. This decrease in carbon dioxide eventually results in a cooler climate. In this way, the weathering at the surface of a feldspar grain is linked to the causes of global climate change. As more and more carbon dioxide is used up through weathering and the climate cools, weathering decreases again. As weathering decreases, the amount of carbon dioxide in the atmosphere builds up again, and the

climate warms, thus completing the cycle. We consider this system further in Chapters 17 and 23.

The Role of Carbon Dioxide in Weathering The reaction of feldspar with pure water in a laboratory is an extremely slow process. Under laboratory conditions, it would take thousands of years to weather even a small amount of feldspar completely. Thus, this reaction cannot account for the more rapid weathering that we observe widely in nature. If we wanted to, we could speed weathering by adding a strong acid (such as hydrochloric acid) to our solution and dissolve the feldspar in a few days. An acid is a substance that releases hydrogen ions (H^+) to a solution. A strong acid produces abundant hydrogen ions; a weak one, relatively few. The strong tendency of hydrogen ions to combine chemically with other substances makes acids excellent solvents.

On Earth's surface, the most common natural acid—and the one responsible for increasing weathering rates—is carbonic acid (H_2CO_3). This weak acid forms when carbon dioxide (CO_2) gas from the atmosphere dissolves in rainwater:

$$\text{carbon dioxide} + \text{water} \rightarrow \text{carbonic acid}$$
$$CO_2 \qquad H_2O \qquad H_2CO_3$$
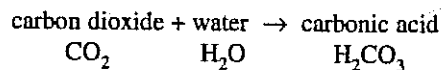
We are familiar with everyday solutions of carbon dioxide in water in the form of carbonated soft drinks. The bottler carbonates the liquid by pumping carbon dioxide into it under pressure. A large quantity of carbon dioxide becomes dissolved in the beverage, making it acidic. As you open the bottle, the pressure drops and the dissolved gas bubbles out of solution, making the solution less acidic. The amount of carbon dioxide dissolved in the liquid decreases as the amount of gaseous carbon dioxide in contact with the water decreases. When the amount of carbon dioxide in the beverage reaches the amount in the atmosphere, no more carbon dioxide bubbles out and the beverage is "flat" and only mildly acidic, like rainwater.

The amount of carbon dioxide dissolved in rainwater is small because the amount of carbon dioxide gas in the atmosphere is small. About 0.03 percent of the molecules in Earth's atmosphere are carbon dioxide. Small as that number is, it makes carbon dioxide the fourth most abundant gas, just behind argon (0.9 percent), oxygen (21 percent), and nitrogen (78 percent). The amount of carbonic acid formed in rainwater is very small, only about 0.0006 g/l. As the burning of oil, gas, and coal increases the amount of carbon dioxide in the atmosphere, the amount of carbonic acid in rain increases slightly.

Most of the acidity of acid rain, however, comes not from carbon dioxide but from sulfur dioxide and nitrogen gases, which react with water to form strong sulfuric and nitric acids, respectively. These acids promote weathering to a greater degree than carbonic acid does. Volcanoes and coastal marshes emit gases of carbon, sulfur, and nitrogen into the atmosphere, but by far the largest source is industrial pollution. (See Chapter 23 for more information about acid rain.)
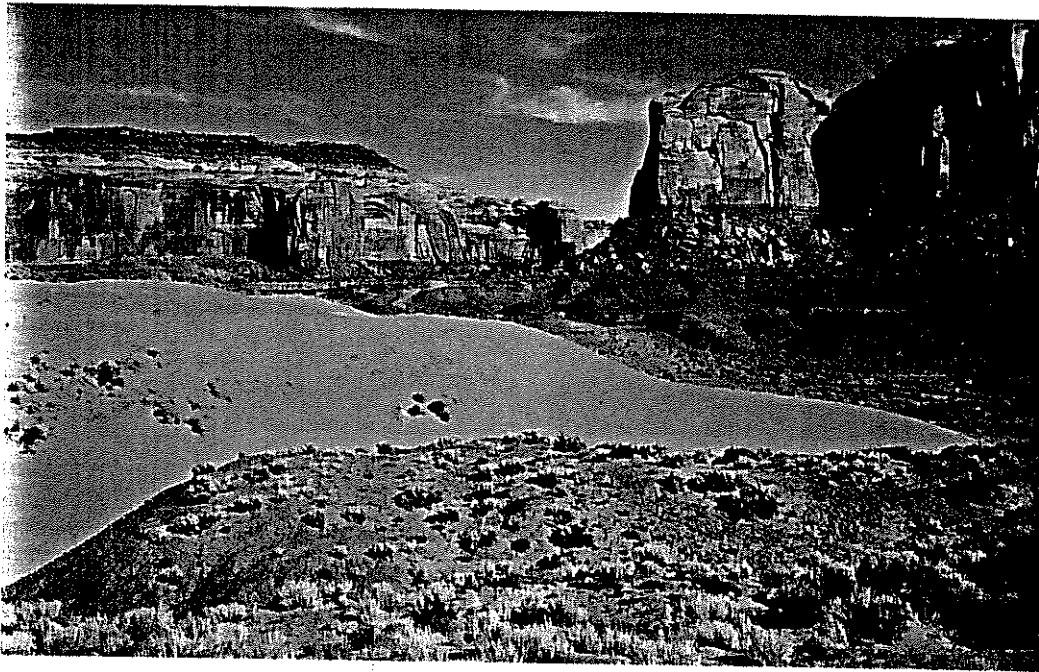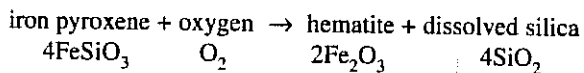
Figure 7.9 Red and brown iron oxides color weathering rocks in Monument Valley, Arizona. [Betty Crowell.]

tates from the solution, forming a solid ferric iron oxide. We are familiar with ferric iron oxide in another form—rusting iron, which is produced when iron metal is exposed to the atmosphere.

We can show this overall weathering reaction of iron-rich minerals by the following equation:

iron pyroxene + oxygen → hematite + dissolved silica
$$4FeSiO_3 \qquad O_2 \qquad 2Fe_2O_3 \qquad 4SiO_2$$

Although the equation does not show it explicitly, water is required for this reaction to proceed.

Iron minerals, which are widespread, weather to the characteristic red and brown colors of oxidized iron (Figure 7.9). Iron oxides are found as coatings and encrustations that color soils and weathered surfaces of iron-containing rocks. The red soils of Georgia and other warm, humid regions are colored by iron oxides. Iron minerals weather so slowly in frigid regions that iron meteorites frozen in the ice of Antarctica are almost entirely unweathered.

Physical Weathering

Now that we have surveyed chemical weathering alone, we can turn to its partner, physical weathering. We can see the workings of physical weathering most clearly by examining its role in arid regions, where chemical weathering is minimal.

Physical Weathering in Arid Regions

Weathered outcrops in arid regions are covered by a rubble of various-sized fragments, from individual mineral grains only a few millimeters in diameter to boulders more than a meter across. The differences in size result from varying degrees of physical weathering and from patterns of breakage of the parent rock. As physical weathering proceeds, the larger particles are cracked and broken into smaller ones. Some of these fragments are broken along planes of weakness in the parent rock (Figure 7.10). Grains of sand form when individual crystals of various minerals, such as quartz, break apart from one another or when fine-grained rocks such as basalt are fragmented.

Although physical weathering is the most common form of weathering in dry regions, even there chemical weathering has prepared the way. Slight chemical alterations of feldspar and other minerals weaken the cohesive forces that hold the crystals in a rock together. As small cracks form and widen, individual quartz or feldspar crystals are freed by a combination of physical and chemical weathering and fall to the ground. Fractures enlarge, and large blocks of rock are separated from an outcrop.

Physical Weathering in All Regions

The chemical weathering that promotes physical weathering is itself promoted by fragmentation, which opens channels through which water and air can penetrate and react with minerals inside a rock. The breaking up of a rock into

which are detailed in Chapters 12 through 16, contribute to the formation of different kinds of landscape (Chapter 17).

Soil: The Residue of Weathering

Not all weathering products are eroded and immediately carried away by streams or other transport agents. On moderate and gentle slopes, plains, and lowlands, a layer of loose, heterogeneous weathered material remains overlying the bedrock. It may include particles of weathered and unweathered parent rock, clay minerals, iron and other metal oxides, and other products of weathering. Engineers and construction workers refer to this entire layer as "soil." Geologists, however, prefer to call this material **regolith,** reserving the term *soil* for the topmost layers, which contain organic matter and can support life. We can easily see the difference between regolith and soil if we consider the regolith found on the Moon. The Moon's regolith is a loose layer of fragmented rocks and dust, but it is quite sterile. It contains little or no organic matter and cannot support life. The organic matter in Earth's soil is **humus,** the remains and waste products of the many plants, animals, and bacteria living in it. Leaf litter contributes significantly to the soil of forests.

Soils vary in color, from the brilliant reds and browns of iron-rich soils to the black of soils rich in organic matter. Soils also vary in texture. Some are full of pebbles and sand; others are composed entirely of clay. Soils are easily eroded, and so they do not form on very steep slopes or where high altitude or frigid climate prevents plant growth.

Soil is such an essential part of our environment and economy that a separate field of study, soil science, developed in the twentieth century. Soil scientists, agronomists, geologists, and engineers study the composition and origin of soils, their suitability for agriculture and construction, and their value as a guide to climate conditions in the past. Many scientists are focusing special attention on ways to combat the serious threat of soil erosion (see Feature 7.1).

Soil Profiles

A road or trench cut through a soil reveals its vertical structure, the soil profile (Figure 7.16). The soil's topmost layer, typically not much more than a meter or two thick, is usually the darkest, containing the highest concentration of organic matter. This topmost layer of soil is called the **A-horizon** (a particular level in a rock section is commonly called a "horizon"). In a thick soil that has formed over a long period of time, the inorganic components of this top layer are mostly clay and insoluble minerals such as quartz. Soluble minerals have been leached from this layer. Beneath the topmost section is the **B-horizon,** where organic matter is sparse. In this layer, soluble minerals and iron oxides have accumulated in small pods, lenses, and coatings. The lowest layer,

the **C-horizon,** is slightly altered bedrock, broken and decayed, mixed with clay from chemical weathering. The transition from one horizon to another is usually indistinct.
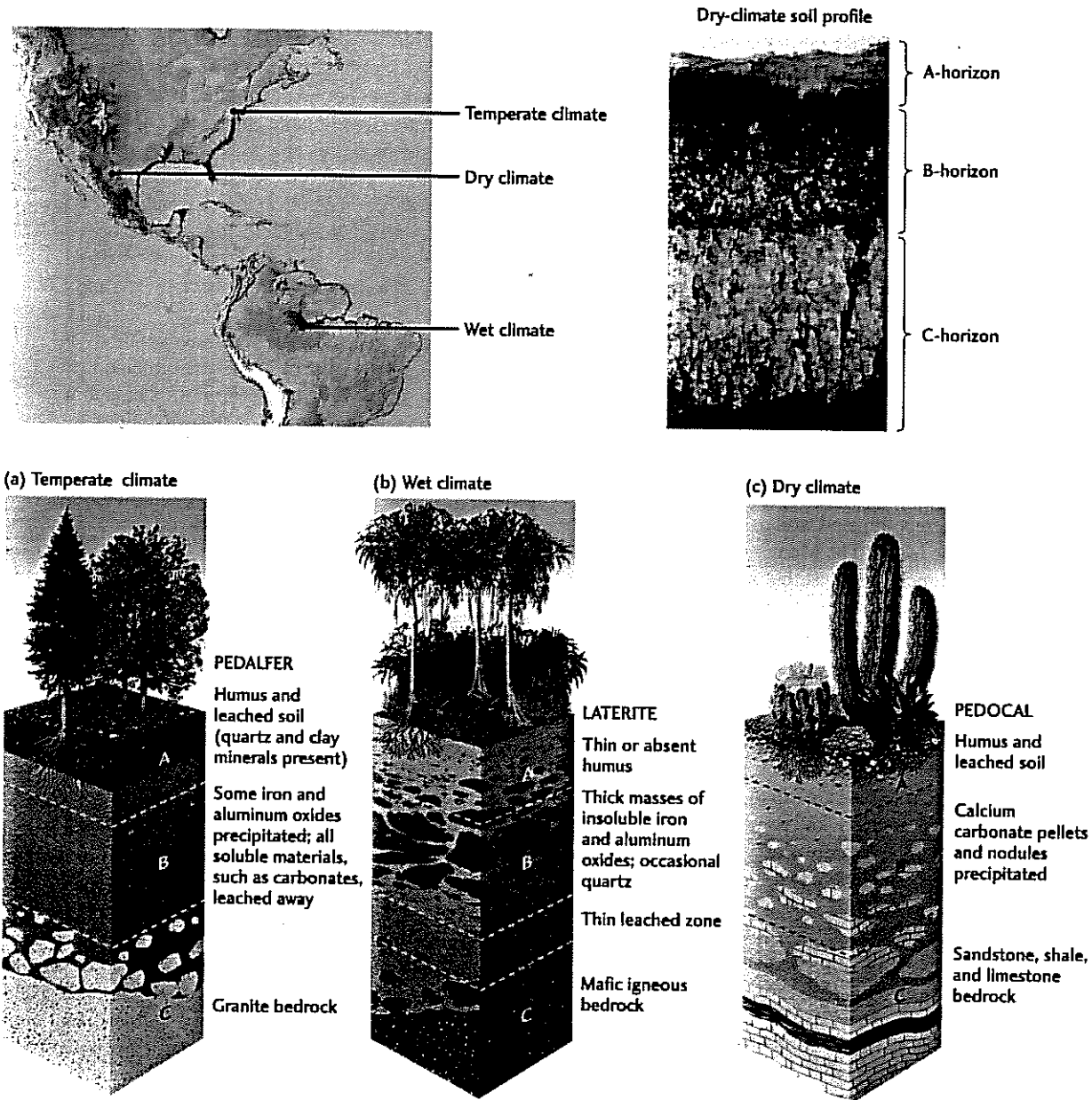
Soils are also described as either residual or transported. Residual soils evolve in one place from bedrock to well-developed soil horizons. Most soils are residual, and they form faster and become thicker where weathering is intense. Even under intense weathering, the A-horizon may take thousands of years to develop to the point at which it can support crops. Soil forms slowly because the most active chemical weathering occurs only during short intervals of rain. During dry periods, the reactions continue, but very slowly and only when some moisture remains in the soil. When soil dries out fully between rains, chemical weathering stops almost completely.

Transported soils may accumulate in some limited areas of lowlands after being eroded from surrounding slopes and carried downhill. They can often be recognized because they are more similar in texture and composition to soils than to ordinary sediments. However, they may be confused with ordinary sediment laid down by rivers, wind, and ice. In some cases, parts of the original upland soil profile are preserved. These soils owe their thickness to deposition rather than to weathering in place. Transported soils are common.

Climate, Time, and Soil Groups

Climate strongly affects weathering and therefore has a great influence on the characteristics of the soil formed on any given parent rock. The characteristics of a soil in a warm and humid region, for instance, differ from those of soils in arid and temperate regions. Soil scientists have mapped soil characteristics over much of the world, with the hope of preventing soil erosion and fostering efficient agricultural practices. For our purposes, we can distinguish three major soil groups—pedalfers, laterites, and pedocals—on the basis of their mineralogy and chemical composition, which also correlate with climate (see Figure 7.16).

Temperate Climates: Pedalfers The characteristics of soils in regions with moderate rainfall and temperatures depend on the climate, the type of parent rock, and the length of time the soil has had to develop and thicken. Intense weathering decreases the influence of the parent rock, as does the length of exposure to weathering. Soil developed after a relatively short time on a granite bedrock in a climate with moderate temperatures and humidity may differ greatly from soil formed on limestone under the same conditions, because the influence of the parent rock remains strong. The soil on the granite may contain remnants of the silicate minerals and be dominated by the clay minerals forming from feldspar, a chief constituent of the parent rock. The soil on the limestone may have a few remnants of calcium carbonate, but most of the limestone fragments will have dissolved. The clay minerals will be mainly those

Dry-climate soil profile

A-horizon

B-horizon

C-horizon

Temperate climate

Dry climate

Wet climate

(a) Temperate climate

PEDALFER

Humus and leached soil (quartz and clay minerals present)

Some iron and aluminum oxides precipitated; all soluble materials, such as carbonates, leached away

Granite bedrock

(b) Wet climate

LATERITE

Thin or absent humus

Thick masses of insoluble iron and aluminum oxides; occasional quartz

Thin leached zone

Mafic igneous bedrock

(c) Dry climate

PEDOCAL

Humus and leached soil

Calcium carbonate pellets and nodules precipitated

Sandstone, shale, and limestone bedrock

Figure 7.16 Soil profile. The thickness of the soil profile depends on the climate, the length of time during which the soil has been forming, and the composition of the parent rock. The transition from one horizon to another is usually indistinct. (a) Pedalfer soil profile developed on granite in a region of high rainfall. The only mineral materials in the upper parts of the soil profile are iron and aluminum oxides and silicates such as quartz and clay minerals, all of which are very insoluble. Calcium carbonate is absent. (b) Laterite soil profile developed on a mafic igneous rock in a tropical region. In the upper zone, only the most insoluble precipitated iron and similar oxides remain, plus occasional quartz. All soluble materials, including even relatively insoluble silica, are leached; thus the whole soil profile may be considered to be an A-horizon directly overlying a C-horizon. (c) Pedocal soil profile developed on sedimentary bedrock in a region of low rainfall. The A-horizon is leached; the B-horizon is enriched in calcium carbonate precipitated by evaporating soil waters. [Photo courtesy of Department of Primary Industries, Victoria, Australia.]
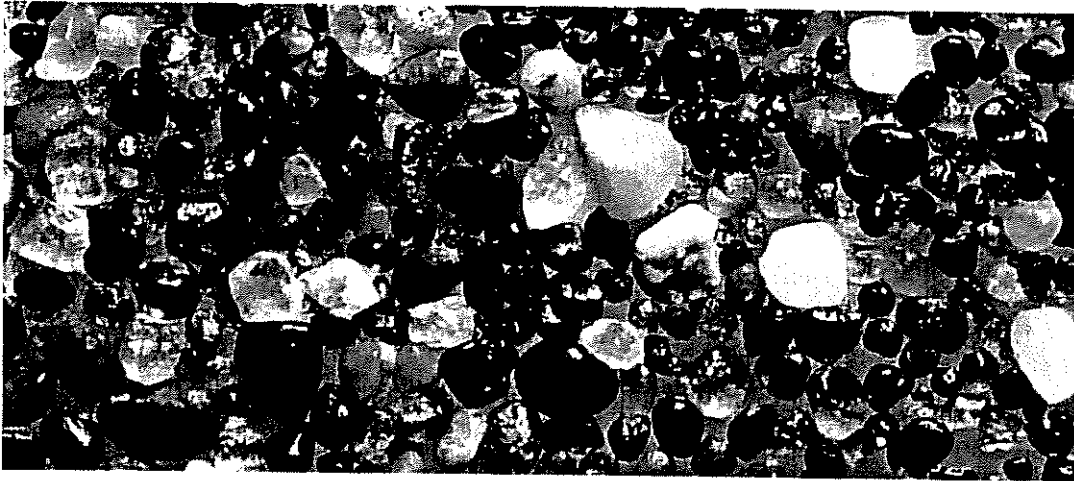
Figure 7.17 Sand is a sediment composed of grains of various minerals broken down from various parent rocks in the course of chemical and physical weathering. [Rex Elliott.]

In this way, a granite rock is transformed into the raw material of sediment and the salts of the sea. The weathered products are eventually carried away from the weathering site by wind, water, and ice. Ultimately they are deposited as various kinds of sediment, such as the sand, silt, and mud found in river valleys and the limestones and evaporites of the world's oceans. As the rock cycle proceeds, this sediment is buried by additional deposits and slowly turns into sedimentary rock, the subject of Chapter 8.

SUMMARY

What is weathering and how is it geologically controlled? Rocks are broken down at Earth's surface by chemical weathering—the chemical alteration or dissolution of a mineral—and by physical weathering—the fragmentation of rocks by mechanical processes. Erosion wears away the land and transports the products of weathering, the raw material of sediment. The nature of the parent rock affects weathering because various minerals weather at different rates. Climate strongly affects weathering: warmth and heavy rainfall speed weathering; cold and dryness slow it down. The presence of soil accelerates weathering by providing moisture and an acidic environment, which promote chemical weathering and the growth of plant roots that aid physical weathering. All else being equal, the longer a rock weathers, the more completely it breaks down.

How does chemical weathering work? Potassium feldspar $(KAlSi_3O_8)$, or orthoclase, weathers in much the same way as do many other kinds of silicate minerals. Feldspar weath-ers in the presence of water by hydrolysis. In the chemical reaction of orthoclase with water, potassium (K) and silica (SiO_2) are lost to the water solution and the solid feldspar changes into the clay mineral kaolinite, $Al_2Si_2O_5(OH)_4$. Carbon dioxide (CO_2) dissolved in water promotes chemical weathering by providing acid in the form of carbonic acid (H_2CO_3). Water in the ground and streams on the surface carry away dissolved ions and silica. Iron (Fe), which is found in ferrous form in many silicates, weathers by oxidation, producing ferric iron oxides in the process. Carbonates weather by dissolving completely, leaving no residue. These processes operate at varying rates, depending on the chemical stability of minerals under weathering conditions.

What are the processes of physical weathering? Physical weathering breaks rocks into fragments, either along crystal boundaries or along joints in rock masses. Physical weathering is promoted by chemical weathering, which weakens grain boundaries; by frost wedging, crystallization of minerals, and burrowing and tunneling by organisms and tree roots, all of which expand cracks; by fires; and perhaps by alternating extremes of heat and cold. Patterns of breakage such as exfoliation and spheroidal weathering result from interactions between chemical and physical weathering processes.

How do soils form as products of weathering? Soil is a mixture of clay minerals, weathered rock particles, and organic matter that forms as organisms interact with weathering rock and water. Weathering is controlled by climate and the activity of organisms, so soils form faster in warm, humid climates than in cold, dry ones. Young soils are affected by the composition of the parent rock, but older soils

Sediments and Sedimentary Rocks

"A man should examine for himself the great piles of superimposed strata, and watch the rivulets bringing down mud, and the waves wearing away the sea-cliffs, in order to comprehend something about the duration of past time, the monuments of which we see all around us."

CHARLES DARWIN

Much of Earth's surface, including its seafloor, is covered with sediments. These layers of loose particles have diverse origins. Most sediments are created by weathering of the continents. Some are the remains of organisms that secreted mineral shells. Yet others consist of inorganic crystals that precipitated when dissolved chemicals in oceans and lakes combined to form new minerals.

Sedimentary rocks were once sediments, and so they are records of the conditions at Earth's surface when and where the sediments were deposited. Geologists can work backward to infer the sources of the sediments from which the rocks were formed and the kinds of places in which the sediments were originally deposited. For example, the top of Mount Everest is composed of fossiliferous limestones. This evidence indicates that long before Mount Everest was uplifted, this point—now the highest in the world—was part of the floor of an ocean.

The kind of analysis used to make inferences about rock formations at the top of Mount Everest applies just as well to ancient shorelines, mountains, plains, deserts, and swamps in other regions. In one area, for example, sandstone may record an earlier time when beach sands accumulated along a shoreline that no longer exists. In a bordering area, carbonate reefs may have been laid down along the perimeter of a tropical island. Beyond, there may have been a nearshore area in which the sediments were shallow marine carbonate muds that later became thin-bedded limestone. In reconstructing such environments, we can map the continents and oceans of long ago.

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Further inferences are also possible. We can work backward to understand former plate tectonic settings and movements by studying sedimentary rocks, which reveal their origin in volcanic arcs, rift valleys, or collisional mountains. In some cases, where the components of sediments and sedimentary rocks are derived from the weathering of preexisting rocks, we can form hypotheses about the ancient climate and weathering regime. We can also use sedimentary rocks formed by precipitation from seawater to read the history of change of Earth's climate and seawater chemistry.

The study of sediments and sedimentary rocks has great practical value as well. Oil and gas, our most valuable sources of energy, are found in these rocks. So is much of the uranium used for nuclear power. Coal, which is a distinct type of sedimentary rock, is also used to generate energy. Phosphate rock used for fertilizer is sedimentary, as is much of the world's iron ore. Knowing how these kinds of sediments form helps us to find and use these limited resources.

Finally, because virtually all sedimentary processes take place at or near Earth's surface where we humans live, they provide a background for our understanding of environmental problems. We once studied sedimentary rocks primarily to understand how to exploit the natural resources just mentioned. Increasingly, however, we study these rocks to improve our understanding of Earth's environment.
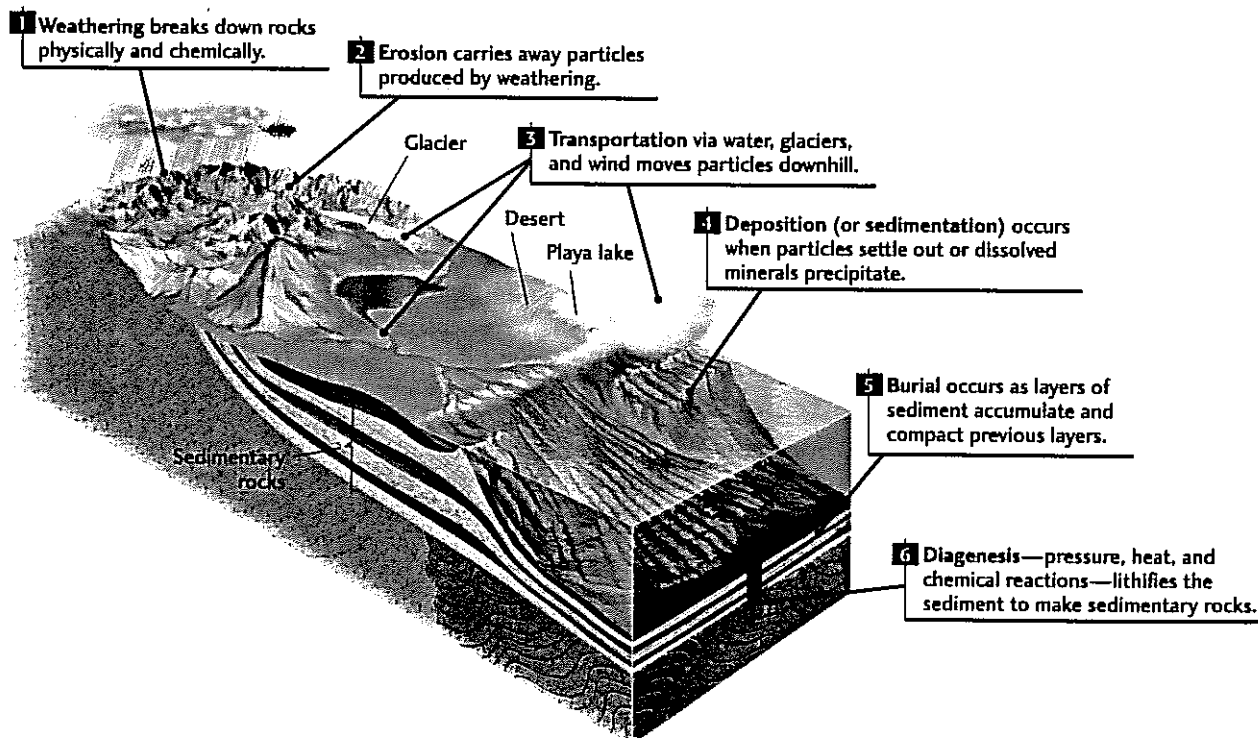
In this chapter, we will see how geologic processes such as weathering, transportation, sedimentation, and diagenesis produce sediments and sedimentary rocks. We will describe the compositions, textures, and structures of sediments and sedimentary rocks and examine how they correlate with the kinds of environments in which the sediments and rocks are laid down. Throughout the chapter, we will apply our understanding of sediment origins to the study of human environmental problems and to the exploration for energy and mineral resources.

Sedimentary Rocks and the Rock Cycle

Sediments, and the sedimentary rocks formed from them, are produced during the surface stage of the rock cycle (discussed in Chapter 4). In other words, they form after rocks have been brought up from the interior by tectonics and before they are returned to the interior by burial. The processes that make up the sedimentary stages of the rock cycle are reviewed in **Figure 8.1**.

• *Weathering* Physical weathering fractures rocks; chemical weathering converts minerals and rocks into altered solids, solutions, and precipitates.

1 Weathering breaks down rocks physically and chemically.

2 Erosion carries away particles produced by weathering.

Glacier

3 Transportation via water, glaciers, and wind moves particles downhill.

Desert

Playa lake

4 Deposition (or sedimentation) occurs when particles settle out or dissolved minerals precipitate.

5 Burial occurs as layers of sediment accumulate and compact previous layers.

Sedimentary rock

6 Diagenesis—pressure, heat, and chemical reactions—lithifies the sediment to make sedimentary rocks.

Figure 8.1 The sedimentary stages of the rock cycle comprise several overlapping processes: physical and chemical weathering; erosion; transportation; deposition; burial; and diagenesis.

• *Erosion* Erosion mobilizes the particles produced by weathering.

• *Transportation* Currents of wind and water and the moving ice of glaciers transport particles to new locations downhill or downstream.

• *Deposition* (also called *sedimentation*) Sedimentary particles settle out as winds die down, water currents slow, or glacier edges melt. These particles form layers of sediment on land or under the sea. In the ocean or in land aquatic environments, chemical precipitates form and are deposited, and the shells of dead organisms are broken up and deposited.

• *Burial* As layers of sediment accumulate, older deposited material is compacted and eventually buried in Earth's crust.

• *Diagenesis* Diagenesis refers to the physical and chemical changes—including pressure, heat, and chemical reactions—by which buried sediments are lithified and acquire a new identity as sedimentary rocks.

Sediments in the Earth System

We can view the sedimentary stages of the rock cycle as interactions between the plate tectonic and climate geosystems that govern Earth's surface and shallow crust. For example, weathering and erosion result from interactions of the solid crust with the atmosphere and oceans. In a typical interaction, weathering and transportation might increase if rainfall increases. Transportation brings the materials to deposition sites. As deposits accumulate, they are buried by later arrivals of sediment. As sediments are buried, they begin to undergo diagenesis. In this chapter and succeeding ones, we will see how specific processes, such as stream flow, play roles in these system interactions.

Weathering and Erosion Yield the Raw Materials: Particles and Dissolved Substances

As we saw in Chapter 7, chemical weathering and mechanical fragmentation of rock at the surface make both solid and dissolved products, and erosion carries away these materials. The end products are grouped either as clastic sediments or as chemical and biochemical sediments.

Clastic Sediments Clastic particles are physically transported rock fragments produced by the weathering of pre-existing rocks. **Clastic sediments** are accumulations of clastic particles. The particles range in size from boulders to pebbles to particles of sand, silt, and clay. They also vary widely in shape. Natural breakage along joints, bedding planes, and other fractures in the parent rock determines the shapes of boulders, cobbles, and pebbles. Sand grains tend to inherit their shapes from the individual crystals formerly interlocked in the parent rock.

Clastic sediments are sometimes called *siliciclastic* because they are produced by the weathering of rocks composed largely of silicate minerals. The mixture of minerals in clastic sediments varies. Minerals such as quartz are resistant to weathering and thus are found unaltered in clastic sediments. There may be partly altered fragments of minerals, such as feldspar, that are less resistant to weathering and so less stable. Still other minerals in clastic sediments, such as clay minerals, may be newly formed. Varying intensities of weathering can produce different sets of minerals in sediments derived from the same parent rock. Where weathering is intense, the sediment will contain only clastic particles made of chemically stable minerals, mixed with clay minerals. Where weathering is slight, many minerals that are unstable under surface conditions will survive as clastic particles. Table 8.1 shows three possible sets of minerals in a typical granite outcrop.

Chemical and Biochemical Sediments The dissolved products of weathering are ions or molecules in the waters of soils, rivers, lakes, and oceans. These dissolved substances are precipitated from water by chemical and biochemical reactions. **Chemical sediments** form at or near their place of deposition, usually from seawater. **Biochemical sediments** comprise the undissolved mineral remains of organisms as well as minerals precipitated by biological processes. We distinguish these two sediment types here for

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Table 8.1   Minerals Remaining in Clastic Sediments Derived from an Average Granite Outcrop Under Varying Intensities of Weathering

INTENSITY OF WEATHERING		
Low	**Medium**	**High**
Quartz	Quartz	Quartz
Feldspar	Feldspar	Clay minerals
Mica	Mica	
Pyroxene	Clay minerals	
Amphibole		
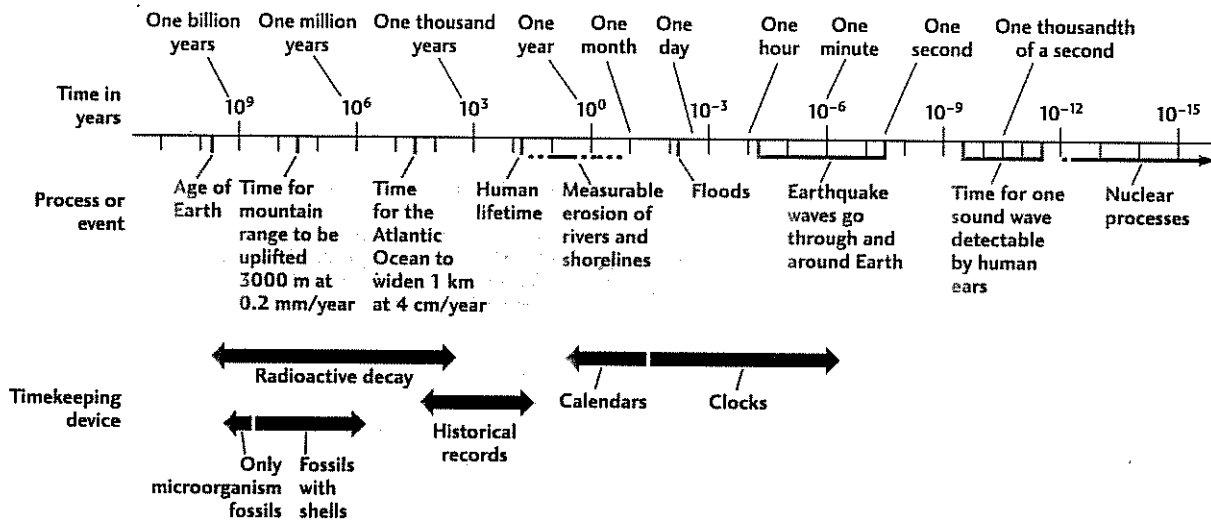
Figure 10.1 The amount of time required for some common processes and events. Times are given as orders of magnitude. The scale is logarithmic; that is, it has equal divisions between successive powers of 10.

atomic nucleus, a rapid chemical reaction. Other scientists perform experiments that last minutes, hours, or days. Geologists, in contrast, deal with Earth processes that unfold over a great range of time periods (**Figure 10.1**). Earthquake tremors last seconds or minutes, whereas the building of a mountain chain takes many millions of years.

We can measure some geological processes directly. A river's floodwaters, for example, rise and fall over a few days. We can even measure the relatively slow movement of glaciers, which may take a year to move a few meters. Other processes, however, such as the erosion of a hillside, are too slow to measure directly. We may rely on historical records to determine the amounts of time required for some of these processes (**Figure 10.2**). Even the oldest historical records extend back only a few thousand years, though, far short of the time span required to measure the very slow geological processes that shape the planet. Our only resource for timing such processes is the rock record. Rocks formed in the past and preserved from erosion serve as Earth's memory, recording geological events, such as glaciations, that lasted many thousands or millions of years.

Geologists of the nineteenth century used their understanding of rock strata and fossils to determine the **relative ages** of sedimentary rock layers—how old they are in relation to one another. These early researchers could then put the geological events that created these rock formations into chronological order. Today, geologists use the physics of radioactive decay to determine a rock's **isotopic age,** often called its absolute age—the actual number of years that have passed since the rock formed.

The geologists who worked out the geologic time scale did more than merely date rocks. They started a revolution in the way we think about time, our planet, and even our-

selves. They discovered that Earth is far older than anyone had previously imagined. Contrary to the common beliefs of earlier times, Earth's surface and interior have been changed and shaped repeatedly by the same geological processes that are at work today. They found, too, that not only the planet but its inhabitants have evolved over time. And humans, they discovered, account for only the briefest moment of Earth's long history.

Timing the Earth System

Our understanding of how Earth systems behaved in the past depends on measurements of the rates of geological processes and their history. For example, we know that carbon dioxide plays an important role in global climate. We can measure carbon dioxide in the atmospheres of the past by drilling ice cores through the Antarctic and Greenland ice caps and analyzing how much carbon dioxide was trapped in the ice as it accumulated over time from countless snowstorms. It is important to know exactly *when* carbon dioxide was high and *when* it was low. So the dating of ice cores has become an important aspect of the dating of climatic variations (see Feature 16.1 in Chapter 16).

Similarly, the history of specific chemical elements—some of great environmental significance, such as lead—can be traced by determining the amount of that element in marine and nonmarine sediments at various times. Thus we can deduce differences in the cycle of lead throughout the past thousands of years. From these differences, we can infer how much lead, a poisonous element, was in the diet of earlier peoples, from Peruvian Incas to Romans at the height of the Roman Empire.
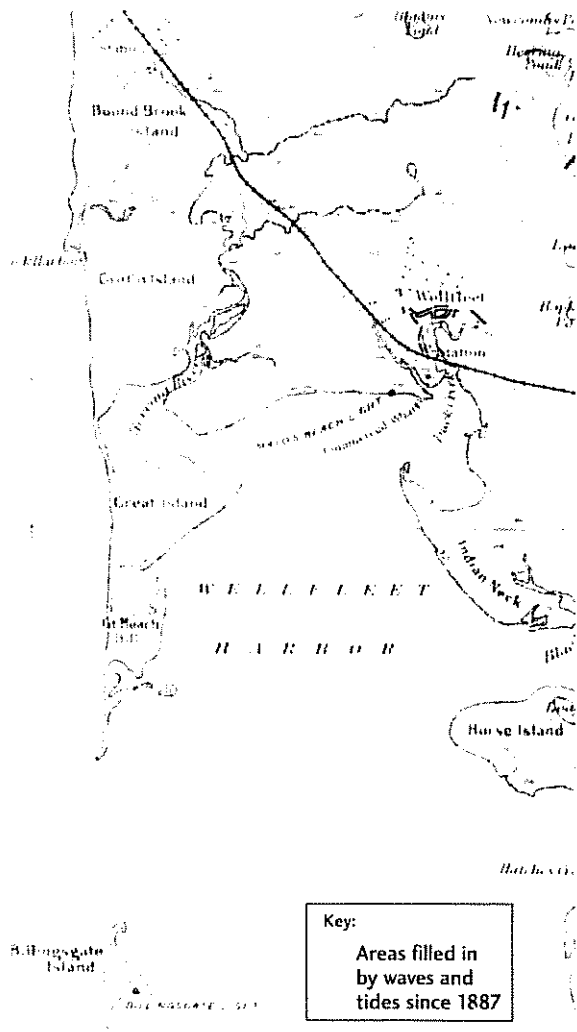
Key:

Areas filled in
by waves and
tides since 1887

Figure 10.2 Historical records such as old maps and land surveys are useful for the measurement of some geological processes. Since 1887, when this map was being prepared, sand, silt, and mud (brown areas added to original in 1988) brought in by waves and tides have filled in areas of tideland near Wellfleet Bay on Cape Cod in Massachusetts.

The only record we have of past geologic events is the incomplete one found in the rocks that have survived erosion or subduction. Because only seafloor younger than 200 million years has survived subduction, we must focus on the continents to find the old rocks that provide the evidence for most of Earth's history. Some of the methods of reading the rock record include interpreting faults and other structures; finding evidence for uplift and erosion; deducing the envi-

ronment in which sediments were deposited and reconstructing the original condition of rocks that have been deformed or metamorphosed.

Geologists of the nineteenth century built a geologic time scale from the time and space relationships of rocks exposed at the surface or in drill holes. They began, as we do here, with evidence from **stratigraphy**—the description, correlation, and classification of strata in sedimentary rocks. It is important to note that for very recent events in geologic history—for example, episodes of climate change on time scales of hundreds to thousands of years—other stratified materials are used to supplement rocks in providing a calendar of past events on Earth. These materials include tree rings, cores of glacial ice from Antarctica and Greenland, and unlithified sediments on the seafloor. Sedimentary rocks, however, are still the most important stratified materials used to decipher the vast depths of geologic history.

The Stratigraphic Record

Stratification—the layering that is the hallmark of sedimentary rocks—is basic to two simple principles used to interpret geologic events from the sedimentary rock record:

1. The **principle of original horizontality** states that sediments are deposited as essentially horizontal beds. Observation of modern marine and nonmarine sediments in a wide variety of environments supports this generalization. (Although cross-bedding, discussed in Chapter 8, is inclined, the overall orientation of cross-bedded units is horizontal.) If we find a sequence of sedimentary rock layers that is folded or tilted, we know that the rocks were deformed by tectonic stresses after their sediments were deposited.

2. The **principle of superposition** states that each layer of sedimentary rock in a tectonically undisturbed sequence is younger than the one beneath it and older than the one above it. Geological common sense tells us that a younger layer cannot slip beneath a layer that has already been deposited. This principle enables us to view a series of layers as a kind of vertical timeline—that is, a partial or complete record of the time elapsed from the deposition of the lowermost bed to the deposition of the uppermost bed (Figure 10.3).

Thus a vertical set of strata, called a **stratigraphic succession,** is a chronological record of the geologic history of a region. The corresponding timeline based on the sequence is called the **geologic time** spanned by the sequence. (The term *geologic time* is also used to refer to the entire span of time since the Earth formed.) Stratigraphic successions differ from sedimentary sequences, which were discussed in Chapter 8. Sedimentary sequences are vertical changes of lithology in sediments deposited in one environment of sedimentation. A stratigraphic succession is defined more broadly and includes a wide variety of beds of different origins. Whereas the emphasis in sedimentary sequences is the nature of successive sediment types, the emphasis in a
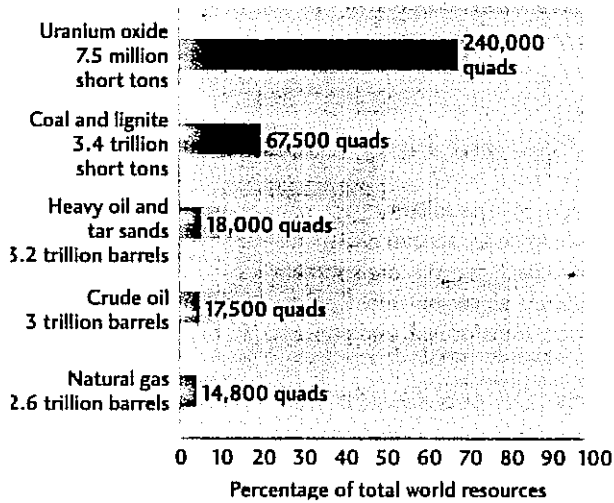
Figure 22.2 A rough estimate of total remaining
nonrenewable world energy resources amounts to about
360,000 quads. Amounts are given in conventional units of
weight (short tons), volume (barrels), and energy content
(quads). A short ton is 2000 pounds, or 907.20 kg; a barrel of
oil is 42 gallons. Coal and lignite resources, for example,
amount to 3.4 trillion short tons, equivalent to 67,500 quads,
or 19 percent of total energy resources. [World Energy Council.]

the nonrenewable sources of energy poses a serious threat to
the environment, and continued use of the ones we depend
on most—coal, oil, and natural gas—will likely trigger
global climate change, with possibly dire consequences.

Energy Use

As the world industrialized, the demand for energy in-
creased and the types of energy used changed. The indus-

trial revolution of the eighteenth and nineteenth centuries
was powered by the energy from coal—in Britain, from the
coalfields of England and Wales; in continental Europe,
from the coal basins of western Germany and bordering
countries; and in North America, from the Appalachian
coalfields of Pennsylvania and West Virginia. As industrial-
ization expanded, so did the hunger for coal. Geological
exploration for this fuel spread over much of the world as
coal use climbed.

Half a century after the first oil well in America was
drilled in 1859, oil and natural gas were beginning to dis-
place coal as the fuels of choice. Not only did they burn
more cleanly, producing no ash, but they could be trans-
ported by pipeline as well as by rail and ship.

In the last quarter of the twentieth century, nuclear
energy was introduced, with expectations that it would pro-
vide a large, new, low-cost, environmentally benign source
of energy. These expectations were not realized, however,
because safety concerns, the inability to dispose of nuclear
wastes, and the escalating costs of stringent safety measures
slowed the construction of nuclear power plants.

The advanced nations depend primarily on oil, coal,
natural gas, and some nuclear energy. In poorer countries,
wood is an important source of fuel. Figure 22.3 summa-
rizes the history of energy use in the United States. In 1850,
coal accounted for less than 5 percent of U.S. energy use.
Today, oil, coal, and natural gas supply almost 90 percent of
the energy used in the United States. Consumption of this
supply is distributed among residential and commercial
uses (35.8 percent), industrial use (37.0 percent), and trans-
portation (27.1 percent). Approximately half of the energy
produced is lost in distribution and inefficient use.

Fossil Fuels

A century and a half ago, most of the energy used in the
United States came from the burning of wood. A wood fire,
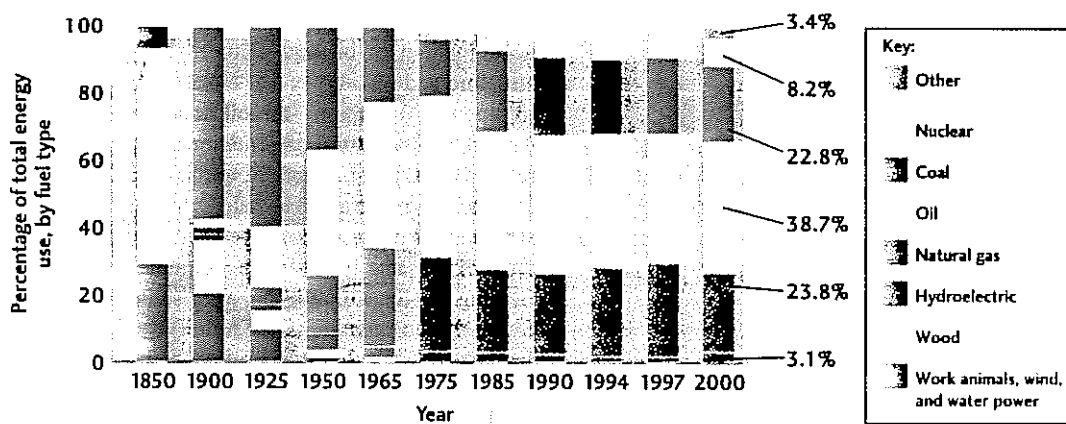
Figure 22.3 Percentages of various types of energy used in the United States from 1850 to
2000. Shown as "Other" for the year 2000 is about 3.4 percent distributed over wood,
geothermal, solar, biomass, wind, and other types of energy. [U.S. Energy Information Agency, 2001.]
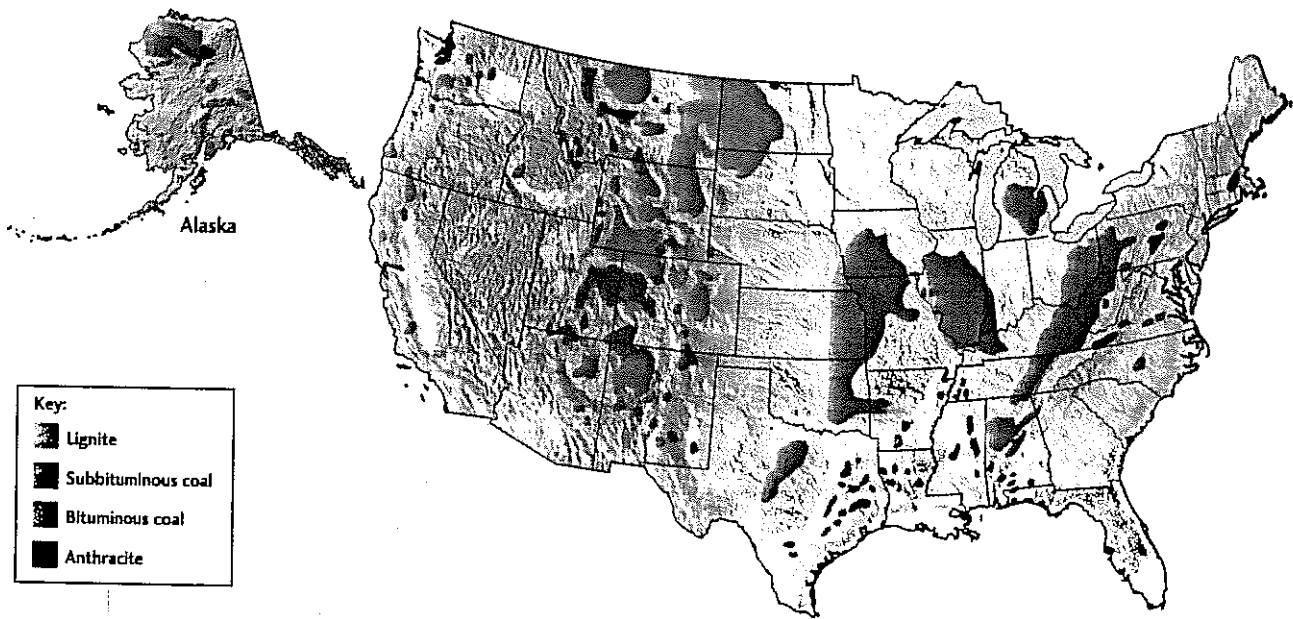
Figure 22.9 Coalfields of the United States. [U.S. Bureau of Mines.]

accumulates and gradually turns into peat, a porous brown mass of organic matter in which twigs, roots, and other plant parts can still be recognized (Figure 22.8). The accumulation of peat in an oxygen-poor environment can be seen in modern swamps and peat bogs. When dried, peat burns readily because it is 50 percent carbon.

Over time, with continued burial, the peat is compressed and heated. Chemical transformations increase the peat's already high carbon content, and it becomes *lignite,* a very soft, brownish-black coal-like material containing about 70 percent carbon. The higher temperatures and structural deformation that accompany greater depths of burial may metamorphose the lignite into *subbituminous* and *bituminous coal,* or soft coal, and ultimately into *anthracite,* or hard coal. The greater the metamorphism, the harder and brighter the coal and the higher its carbon content, which increases its heat value. Anthracite is more than 90 percent carbon.

Coal Resources

According to some estimates (see Figure 22.2), the amount of coal remaining in the world is about 3.4 trillion short tons. The leading producers are the United States (Figure 22.9), the former Soviet Union, and China, which together hold about 85 percent of the world's coal resources (former Soviet Union, 50 percent; China, 20 percent; United States, 15 percent). Domestic coal resources in the United States would last for a few hundred years at current rates of use— about a billion tons a year. Coal has supplied an increasing proportion of U.S. energy needs since 1975, when the price

of oil began to rise, and it currently accounts for about 23 percent of the energy consumed.

Environmental Costs of Coal

There are serious problems with the recovery and use of coal that make it less desirable than oil or gas, whether the coal is burned or converted into synthetic liquid fuel. Much coal contains appreciable amounts of sulfur, which vaporizes during combustion and releases noxious sulfur oxides into the atmosphere. Acid rain, which forms when these gases combine with rainwater, is becoming a severe problem in Canada, Scandinavia, the northeastern United States, and eastern Europe (see Chapter 23). Coal ash is the inorganic residue that remains after coal is burned. It contains metal impurities from the coal, some of which are toxic. Ash can amount to several tons for every 100 tons of coal burned and poses a significant disposal problem. It can escape from smokestacks, posing a health risk to people downwind. Strip mining, the removal of soil and surface sediments to expose coal beds, can ravage the countryside if the land is not restored (Figure 22.10). Underground mining accidents take the lives of miners each year, and many more suffer from black lung, a debilitating inflammation of the lungs caused by the inhalation of coal particles.

Government regulations now require that technologies for the "clean" combustion of coal be phased in. The law mandates the restoration of land ravaged by strip mining and the reduction of danger to miners. But these measures are expensive and add to the cost of coal. This drawback, however, is unlikely to prevent the increased use of this fuel,