

# Rock-Forming Minerals and Rocks

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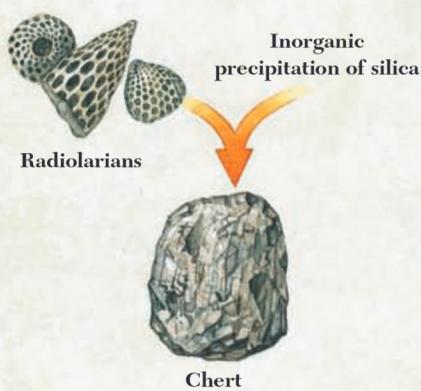
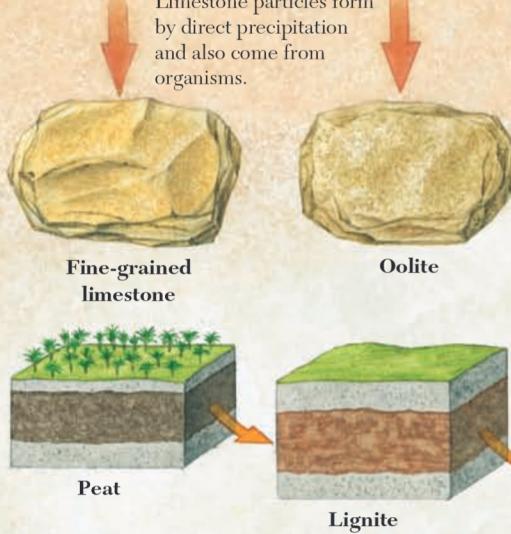
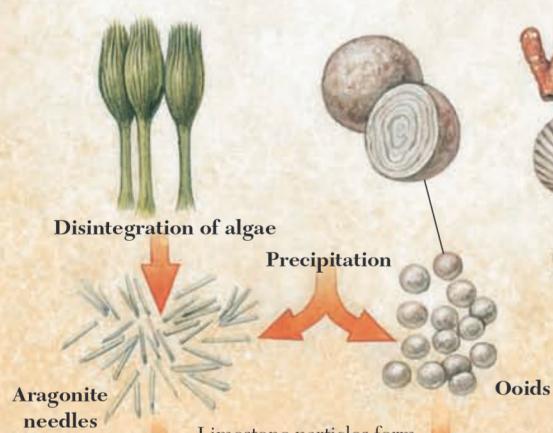
**Stacked pyrite cubes from Peru.**  
Although it is not typically a rock forming mineral, pyrite can be found in most rock types formed throughout Earth's history. Pyrite does play an important role in our understanding of the oxygen content of Earth's atmosphere, which is presented in Chapter 12.  
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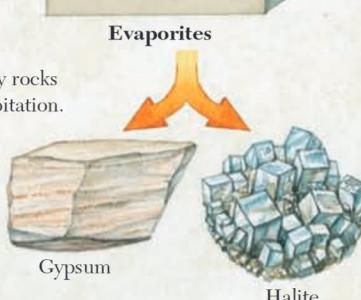
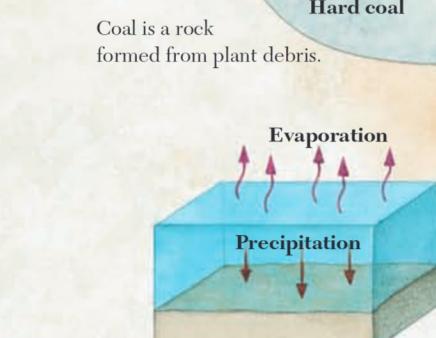
# VISUAL OVERVIEW

## Rocks and Their Origins

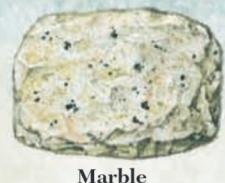
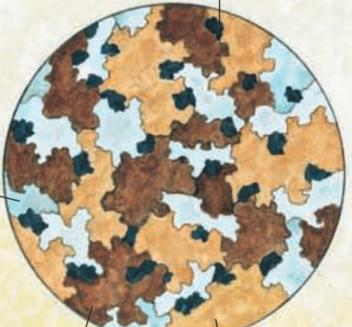
The origins of igneous, sedimentary, and metamorphic rocks are interrelated through the rock cycle.



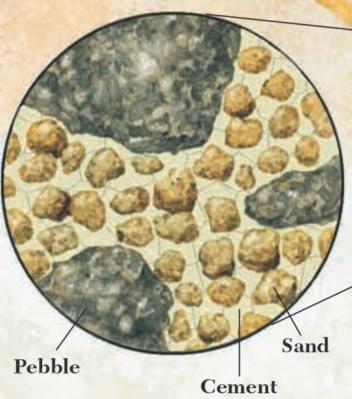
Chemical sedimentary rocks are products of precipitation.

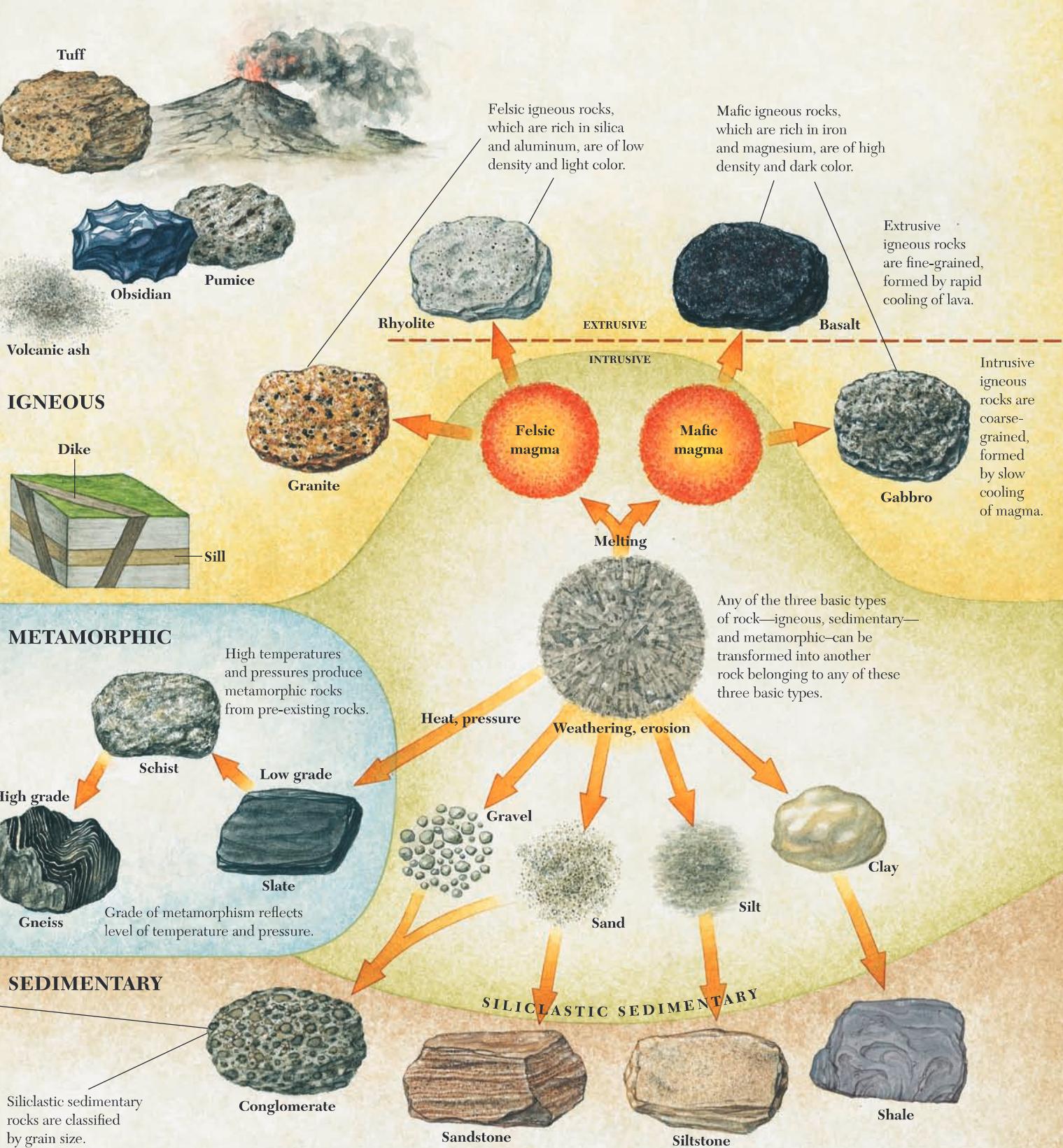


Granite consists of interlocking grains of quartz, feldspar, and mica.

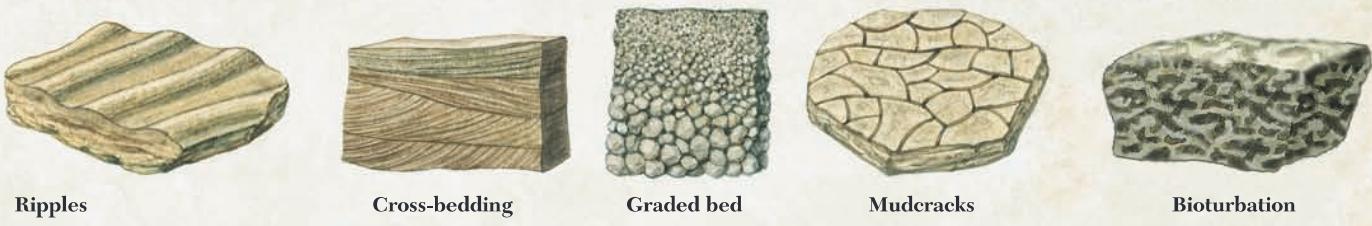


Coal is a rock formed from plant debris.





#### SEDIMENTARY STRUCTURES



**R**ocks, the building blocks of Earth, provide evidence of conditions and events of the geologic past. The composition and configuration of rocks tell stories of mountain-building episodes, changing positions of land and sea, climatic changes, and many other aspects of Earth's history. In this chapter our emphasis is on materials that play prominent roles in the rock cycle: rocks and the minerals that form them.

## The Structure of Minerals

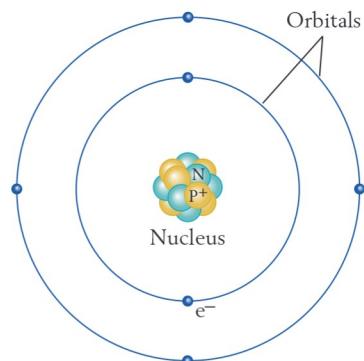
Minerals, as we saw in Chapter 1, are naturally occurring solid elements or compounds. In order to understand the properties of minerals, we need to examine the nature of atoms, the fundamental units of elements. Then we can see how atoms combine to form minerals and how minerals combine to form rocks.

### An element consists of a unique kind of atom

Atoms form elements, which, being indivisible into simpler substances during normal chemical reactions, are the most basic chemically distinct form of matter. At the center of an atom is a nucleus, in which nearly all of the atom's mass resides. Particles called protons and neutrons form the nucleus. Each of these particles has the same mass (1 atomic mass unit); but whereas protons each have a positive charge of +1, neutrons, as their name suggests, have no charge. In orbit around the nucleus are electrons, each of which has a charge of -1. Electrons move about within zones called orbitals or shells around the nucleus (Figure 2-1). Their atomic mass is negligible.

Each chemical element consists of atoms that have a particular number of protons. That number of protons is the element's unique atomic number. Hydrogen, for example, with only one proton, has an atomic number of 1; for oxygen the number is 8, for silicon it is 14, and for iron it is 26 (Figure 2-2). A complete listing of chemical elements and their atomic numbers can be found in the periodic table of the elements.

A neutral, or uncharged, atom has as many electrons as it has protons. Consequently, as the number of protons increases, so does the number of electrons—and the number of orbitals also increases, by steps. The innermost



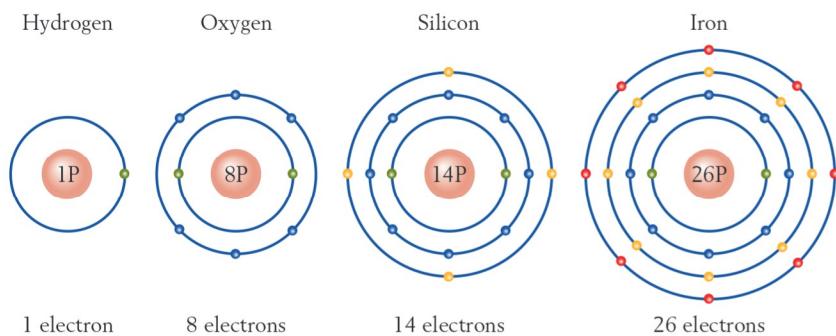
**FIGURE 2-1 Model of a carbon atom.** The central nucleus is surrounded by orbiting electrons ( $e^-$ ). Each proton ( $P^+$ ) has an atomic mass of 1, as does each neutron (N). The mass of an electron is negligible. Protons and electrons have opposite charges.

orbital holds a maximum of two electrons, whereas the second and third shells hold a maximum of eight electrons each. Thus eight of the twenty-six electrons of iron have to occupy a fourth shell.

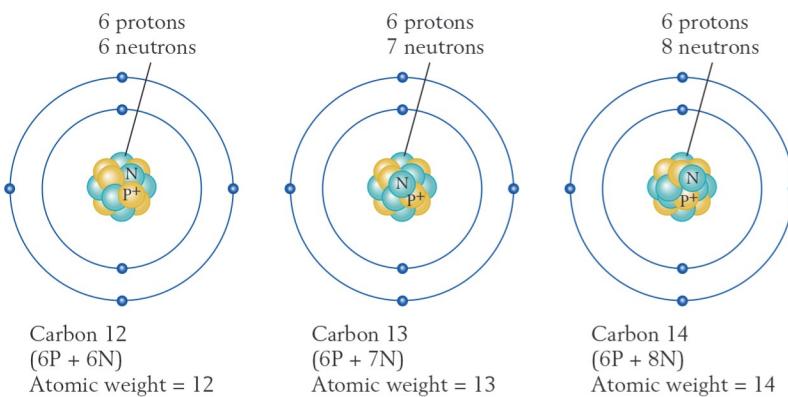
### Isotopes of an element have distinctive atomic weights

Unlike protons, the neutrons of a given element may vary in number. Thus carbon, which has six protons, may have six, seven, or eight neutrons (Figure 2-3). The atomic masses of the protons and neutrons in a carbon atom may therefore add up to 12, 13, or 14. These numbers are termed the *atomic weights* of the different kinds of carbon atoms. Each kind of atom, with its unique atomic weight, is called an **isotope** of its element. The three carbon isotopes are designated carbon 12, carbon 13, and carbon 14.

Isotopes of certain elements have special importance in geology. Some are *radioactive*; that is, their nuclei decay at a constant average rate to form other isotopes, informally termed *daughter isotopes*. The average rate of decay of any radioactive substance can be measured by counting the particles that this decay produces over a given interval of time—for example, by counting the clicks recorded by a nearby Geiger counter that registers the arrival of these particles. Once the average rate of decay



**FIGURE 2-2 Models of several common elements, each with a different number of shells.** Preceding the letter P on each nucleus is the number of protons.



**FIGURE 2-3 Models of three isotopes of carbon.** Carbon 12 and carbon 13 are stable isotopes, whereas carbon 14 is radioactive. ( $P^+$  = proton; N = neutron.)

for a radioactive isotope is known, geologists can calculate the age of a rock containing that isotope by measuring the amount of the parent and daughter isotopes in that rock. The rapid rate at which carbon 14 decays, for example, enables geologists to date carbon-bearing materials, such as wood, that are only a few hundred or a few thousand years old. Most other naturally occurring radioactive isotopes decay much more slowly, and some are widely used to date rocks that are hundreds of millions of years old. The total number of radioactive atoms in Earth is so enormous that their decay releases enough heat to warm the planet significantly (p. 17).

Carbon 12 and carbon 13, the nonradioactive or **stable isotopes** of carbon, are also useful to geologists. As we will see in Chapter 10, the relative abundances of these isotopes within organic matter and fossils in sediments shed light on aspects of Earth's history, including changes in the composition of Earth's atmosphere.

### Chemical reactions produce minerals

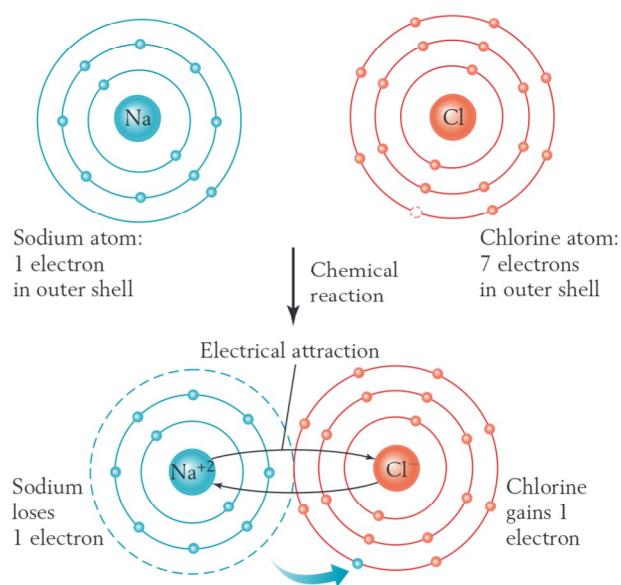
Rocks are made of minerals, but for a mineral to exist, elements must combine by way of a *chemical reaction*, in which two or more atoms interact to form a structure called a *molecule*. An atom of sodium and an atom of chlorine, for example, react chemically to form a molecule of sodium chloride ( $\text{NaCl}$ )—the salt with which we flavor our food. Sodium chloride occurs in nature as the mineral **halite**, informally called rock salt. Halite that was precipitated from ancient seas or salty lakes forms large deposits that we can mine for our use.

Just as an atom is the basic unit of a chemical element, a molecule is the basic unit of a *chemical compound*. Most compounds have very different properties from the elements of which they are formed. Sodium, for example, is a metal, and chlorine is a gas; neither bears any resemblance to sodium chloride, which is a clear, solid compound.

### Chemical reactions create chemical bonds

Molecules form through chemical reactions in which interactions between electrons produce attachments

between atoms known as *chemical bonds*. Molecules tend to be most stable when the outermost shell of each of the atoms forming them contains eight electrons. In one kind of bond, known as an **ionic bond**, one atom loses an electron to another atom. An ionic bond forms in the chemical reaction that produces a molecule of sodium chloride. Elements having one or two electrons in their outer shell tend to lose electrons to other kinds of atoms. Sodium, for example, tends to lose its lone outer electron to chlorine. Chlorine, with an atomic number of 17, has seven electrons in its outermost (third) shell, just one short of the full complement. So if a sodium atom and a chlorine atom approach each other, a chemical reaction can occur in which sodium transfers an electron to chlorine (Figure 2-4). The result is a stable molecule of sodium chloride ( $\text{NaCl}$ ), in which



**FIGURE 2-4 A chemical reaction forms sodium chloride.** Sodium chloride ( $\text{NaCl}$ ) forms by the transfer of the only electron in the outer shell of sodium ( $\text{Na}$ ) to the outer shell of chlorine ( $\text{Cl}$ ). The sodium ion ( $\text{Na}^+$ ) and chloride ion ( $\text{Cl}^-$ ) thus formed are held together by an electrical attraction resulting from their opposite charges.

the outer shells of both the sodium and the chlorine are filled with electrons.

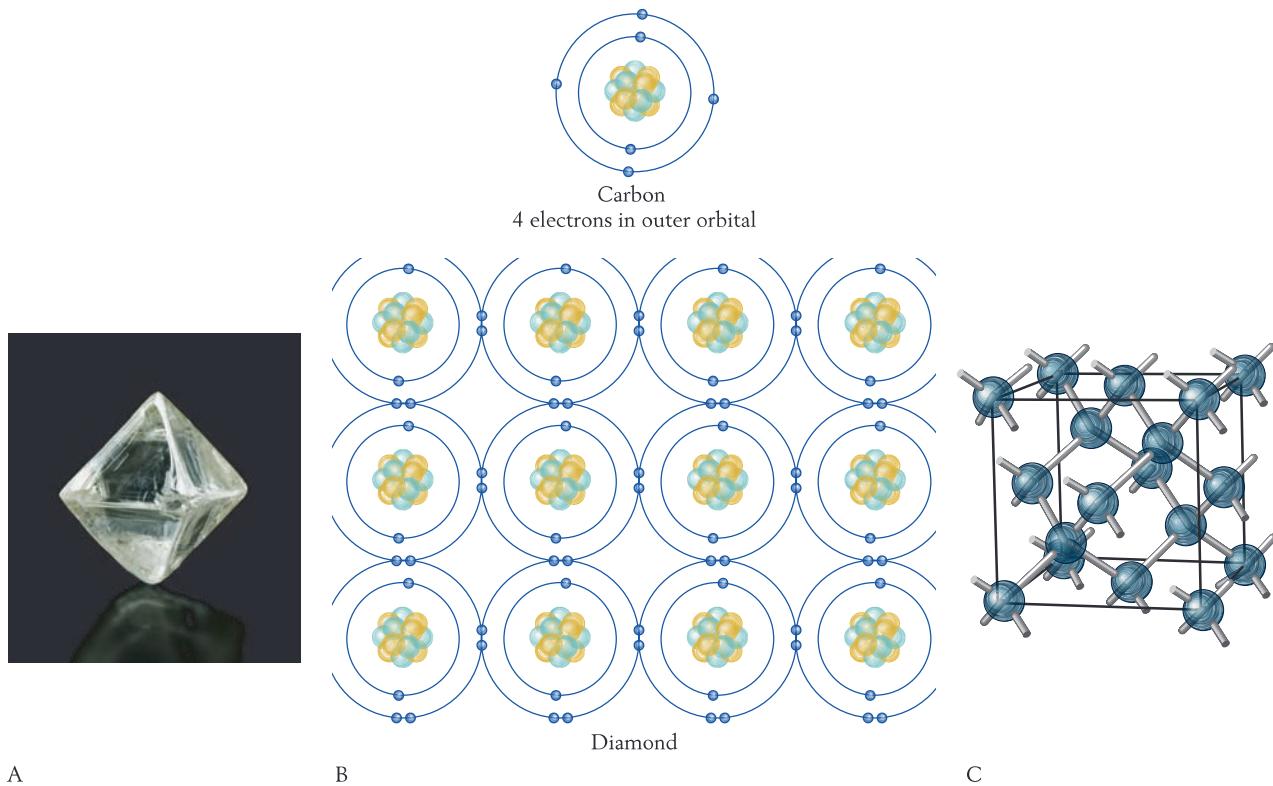
The electron transfer that forms an ionic bond has a substantial effect on the charges of the atoms: each develops a charge imbalance. In the formation of a sodium chloride molecule, for example, the sodium ends up with a net charge of +1 because its loss of an electron leaves its protons outnumbering its electrons by one. In complementary fashion, chlorine develops a charge of -1 by gaining an electron. When an atom becomes charged, it becomes an **ion**. It is the attraction between the positively charged sodium ion and the negatively charged chlorine ion that holds the sodium chloride molecule together.

Ions retain their identity when the compounds they form dissolve in water. When sodium chloride dissolves in water, for example,  $\text{Na}^+$  and  $\text{Cl}^-$  ions are released to form a salt solution. In fact, the accumulation of these and other, less abundant ions is what makes seawater salty. When seawater evaporates in a dry climate,  $\text{Na}^+$  and  $\text{Cl}^-$  can become too highly concentrated to remain in solution. Then  $\text{Na}^+$  and  $\text{Cl}^-$  ions combine, and sodium chloride precipitates. This is how halite deposits form in nature.

Sodium chloride displays the simplest kind of ionic bond, in that a single electron is transferred from sodium to chlorine. Many other kinds of molecules form by the transfer of more than one electron; in fact, many form by the union of three or more ions. Calcium chloride exemplifies both of these complex features; as  $\text{CaCl}_2$ , it consists of a  $\text{Ca}^{2+}$  ion that is attached to two  $\text{Cl}^-$  ions.

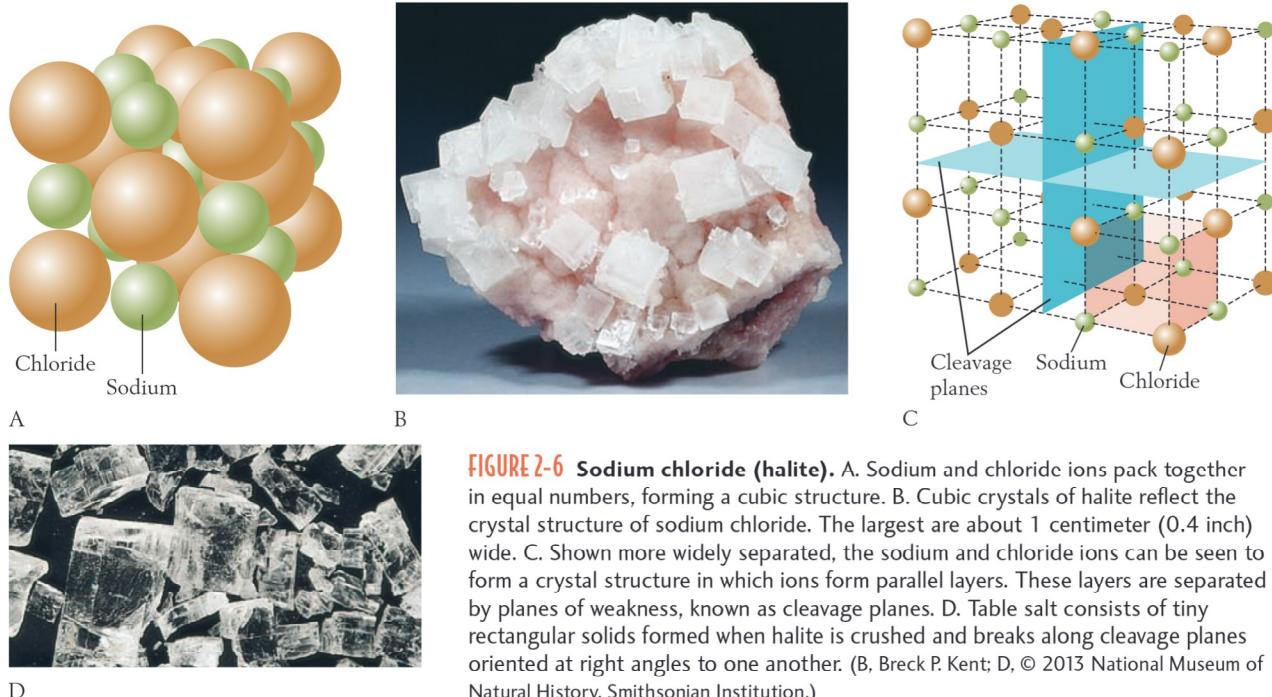
Ions often group together to form *complex ions*. Examples are carbonate ( $\text{CO}_3^{2-}$ ) and sulfate ( $\text{SO}_4^{2-}$ ). Both of these complex ions are abundant in seawater, from which they can precipitate by combining with positive ions to form minerals.

Atoms can also form **covalent bonds**, in which electrons are shared rather than exchanged. The mineral diamond, which is cut to produce gems, consists entirely of carbon atoms united by covalent bonds. An isolated carbon atom, with an atomic number of 6, has four electrons in its outermost (second) shell. When carbon forms diamond, each of the electrons in its outer shell is shared by two atoms (Figure 2-5). Thus the number of electrons in the outer shell of each atom is doubled to the full complement of eight. The result is a stable molecular structure in which every carbon atom shares two electrons with each of four neighboring atoms.



**FIGURE 2-5** Covalent bonding in diamond. A. The mineral diamond is composed of covalently bonded carbon atoms and is often found as octahedral crystals. B. An isolated carbon atom (top) has four electrons in its outer shell. All atoms in diamond (bottom) share the electrons of their outer shell equally so that each atom ends up with a total of eight electrons in its outer

shell instead of four. Because all the electrons of the outer shell are shared, the atoms are strongly bonded, and diamond is very hard. C. The lattice structure of diamond consists of atoms of carbon that are covalently bonded in three dimensions.  
(A, Jeffrey A. Scovil Photography.)



**FIGURE 2-6 Sodium chloride (halite).** A. Sodium and chloride ions pack together in equal numbers, forming a cubic structure. B. Cubic crystals of halite reflect the crystal structure of sodium chloride. The largest are about 1 centimeter (0.4 inch) wide. C. Shown more widely separated, the sodium and chloride ions can be seen to form a crystal structure in which ions form parallel layers. These layers are separated by planes of weakness, known as cleavage planes. D. Table salt consists of tiny rectangular solids formed when halite is crushed and breaks along cleavage planes oriented at right angles to one another. (B, Breck P. Kent; D, © 2013 National Museum of Natural History, Smithsonian Institution.)

### Crystals have three-dimensional molecular structures

For ions to form a stable compound, not only must their ionic charges be in balance, but the ions must have relative diameters that allow them to fit together. In the mineral halite, for example, the small sodium ions fit comfortably between the larger chloride ions (Figure 2-6A).

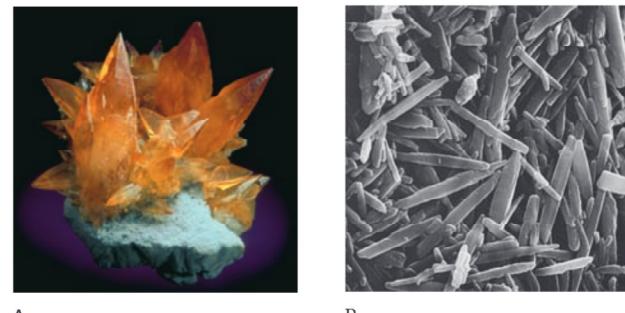
The three-dimensional molecular structure of any crystalline mineral is quite regular, reflecting the relative sizes and numbers of the various kinds of atoms of which it is formed. The external shape of the mineral, in turn, reflects the mineral's internal structure. The structure of halite, for example, consists of equal numbers of  $\text{Na}^+$  and  $\text{Cl}^-$  ions packed closely together in a regular pattern. It is easy to see how this pattern is reflected in the cubic shape of halite crystals (Figure 2-6B).

Two or more minerals can have identical chemical compositions but altogether different crystal structures. **Calcite** and **aragonite**, for example, are common minerals of this kind that both consist of calcium carbonate ( $\text{CaCO}_3$ ). Both of these minerals precipitate from watery solutions in nature, and a variety of organisms secrete one or both to form a skeleton. Calcite forms blocky or tooth-shaped crystals (Figure 2-7A). Aragonite precipitates directly from shallow tropical seas as tiny needle-shaped crystals that accumulate as what is termed **carbonate mud** (Figure 2-7B). This mud and the aragonitic and calcitic skeletons of marine animals are major components of the sediment that hardens to form limestone. Limestones more than a few million years old, however, consist almost entirely of calcite because aragonite is not stable: over time, it is transformed

into calcite. Corals that have lived during the past 250 million years have secreted skeletons of aragonite, but most of these skeletons that have been preserved as fossils were long ago altered to calcite (Figure 2-8). This alteration is often brought about by watery solutions that destroy some of the original features of the coral skeleton.

### Ions of an element can substitute for ions of another similar element

By definition, a particular mineral can have a specified range of chemical compositions (see p. 5). The reason for this latitude in the definition of a mineral is that many



**FIGURE 2-7 Calcite and aragonite, two forms of calcium carbonate ( $\text{CaCO}_3$ ).** A. Calcite crystals, 2–3 centimeters (about 1 inch) in length. B. Electron microscopic view of needle-shaped aragonite crystals; needles like these, which are about 5 micrometers long, form carbonate mud on shallow tropical seafloors. (A, Marvin Dembinsky Photo Associates/Alamy; B, © 2013 National Museum of Natural History, Smithsonian Institution.)



**FIGURE 2-8** **Coral skeletons.** On the left is the cross section of a present-day coral skeleton that consists of aragonite. Its width is about 10 centimeters (4 inches). On the right is a piece of fossil coral of the same species that is only a few hundred thousand years old, but has been largely altered to calcite—a change often brought about by watery solutions. Secondary calcite also fills many of the pores of the original skeleton, obscuring some of the original features. (© 2013 Natural Museum of Natural History, Smithsonian Institution.)

minerals have crystal structures in which a small quantity of a particular ion can substitute for another ion without significantly altering the mineral's physical or chemical properties. Again, calcium carbonate serves to illustrate the point. The strontium ion sometimes substitutes for the calcium ion in the calcium carbonate crystal structure because it is only slightly larger than the calcium ion and has the same charge of +2. Two stable isotopes of strontium occur in Earth's crust and in the ocean. Because strontium ions are present in seawater, some of them find their way into the skeletons of calcium carbonate secreted by marine organisms. The relative amounts of these two isotopes in seawater have changed during Earth's history. Therefore, as Chapter 6 will explain more fully, the isotope ratio of strontium in a fossil can be used to establish the age of the fossil.

## The Properties of Minerals

The molecular structure of a mineral determines many properties that influence its physical and chemical behavior in nature. The chemical properties and mobility of certain chemical elements and the densities of minerals influence whether they are found in Earth's crust or mantle. In addition, their crystal structures and chemical bonding influence how they break and how readily they abrade when they roll about as isolated grains. Table 2-1 shows the chemical and physical properties of the major mineral groups.

### Chemical bonds determine hardness

Molecular bonds within minerals vary greatly in strength. The covalent bond structure of diamond is so strong that diamond is the hardest of all minerals (see Figure 2-5).

Because the powerful covalent bonding that forms diamond can develop only under very high pressures, diamond forms primarily within Earth's mantle and only rarely is elevated to its surface. Very small diamond crystals can also form when a large meteorite strikes Earth, creating enormous pressures at impact.

Graphite, like diamond, is a mineral consisting of pure carbon, but it is much softer. Although graphite consists of sheets of carbon atoms bonded together to form what amounts to one large molecule, these sheets are held together only by the weak mutual attraction of nuclei and electrons of adjacent sheets. The result is that the sheets readily slide past one another and break apart. In fact, graphite is so soft that even paper can abrade it; that is why we can use it as what we call "lead" in our pencils. Graphite forms in Earth's crust, at much lower pressures than are required to produce diamond in the mantle.

### The weight and packing of atoms determine density

**Density** is the mass of a given volume of any substance. The density of a mineral—often expressed in grams per cubic centimeter ( $\text{g}/\text{cm}^3$ )—depends on two things: the atomic weights of the atoms that form the mineral and the degree to which those atoms are packed together. For reference, water has a density of  $1 \text{ g}/\text{cm}^3$ . Iron, with an atomic number of 26, has a greater atomic weight than many other rock-forming elements; thus minerals that contain iron tend to be relatively dense. This is why such minerals are relatively abundant in Earth's mantle. Iron tended to sink deep within Earth early in its history, when the planet was in a hot, liquid state, and it became the dominant element of Earth's core.

We can see how atomic packing increases a mineral's density by comparing diamond and graphite. The tight packing of diamond's carbon atoms, which results from its formation at high pressures deep within Earth, gives it a density of  $3.5 \text{ g}/\text{cm}^3$ ; graphite, which forms under lower pressures, has a density of only  $2.1 \text{ g}/\text{cm}^3$ .

### Fracture patterns reflect crystal structure

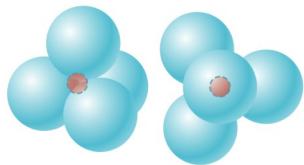
Weak bonding within a mineral's crystal structure can create parallel planes of weakness along which the mineral tends to break. If you observe several grains from a salt shaker through a strong magnifying glass, you will see that many are tiny rectangular blocks. These blocks were formed when larger chunks of halite broke along planes of weakness termed *cleavage planes* (see Figures 2-6C and D).

### Minerals and rocks form under particular physicochemical conditions

Because the compositions and internal structures of minerals reflect the conditions under which the minerals form, these features point to the setting where a

**TABLE 2-1**

Major Mineral Groups				
	Chemical properties	Physical properties	Rock-forming contribution	Comments
Silicates	$\text{SiO}_4^{4-}$ tetrahedra are the basic units	Mostly hard, except for mica and clay minerals; most have a glassy or pearly luster	Dominant mineral group in igneous, sedimentary, and metamorphic rocks	Most crystallize at high temperatures and occur in sediments only as detritus
Carbonates	Positive ions attached to $\text{CO}_3^{2-}$	Soft, light-colored	Mostly sedimentary, but also marble, a metamorphic rock	Include calcite, aragonite, and dolomite
Sulfates	Positive ions attached to $\text{SO}_4^{2-}$	Soft, light-colored, water-soluble	Most rock-forming varieties are sedimentary	Form large sedimentary evaporite deposits, including gypsum and anhydrite
Phosphates	Positive ions attached to $\text{PO}_4^{3-}$	Soft, weakly soluble	Relatively uncommon sedimentary rocks	Form where phosphate is unusually abundant, sometimes from fossil animal bones, fish parts, conodont elements
Halides	Positive ions attached to negative ions of elements such as chlorine (Cl) and fluorine (F)	Soft, light-colored, water-soluble	Most rock-forming varieties are sedimentary	Form large sedimentary deposits, including halite (rock salt); also fluorite
Oxides	Metallic ions combined with oxygen	Soft to hard	Mostly sedimentary, but many varieties are present in igneous and metamorphic rocks	Some, including the iron materials magnetite and hematite, are major ore minerals; aluminum oxides are the ore of aluminum
Sulfides	Metallic ions combined with sulfur	Soft to medium hard, often with a metallic luster	Have a minor role in rock forming; many precipitate from watery fluids	Include pyrite, as well as important ore minerals that form at elevated temperatures
Native elements	Pure forms of elements	Variable; sulfur is yellow, graphite is black, and gold, silver, and copper have metallic luster. Diamond can be any color.	Graphite is usually metamorphosed organic carbon, and the rest are precipitated from hot, watery fluids	Many are valuable or useful to society



Two views of a silica tetrahedron of four oxygens with a silicon hidden in the center

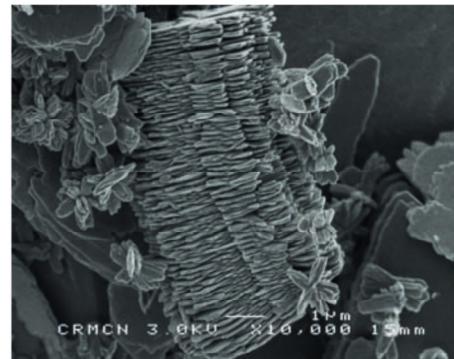
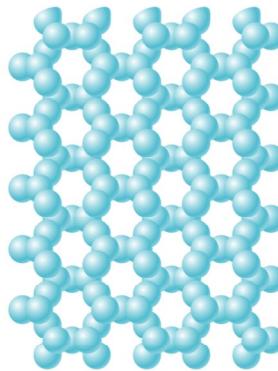


Amphibole (double chain)

In amphiboles and pyroxenes, the silica tetrahedra are assembled into long chains that are bonded together by ions of iron, calcium, or magnesium positioned between them. The iron and magnesium make these minerals dense and often dark.



Mica



Clay

In mica and clay minerals, the silica tetrahedra are more fully connected to form two-dimensional sheets that are bonded together by sheets of aluminum, iron, magnesium, or potassium. Because the bonds between these sheet silicates are weak, micas and clays cleave into thin flakes. Clay minerals are especially weak and almost always occur naturally as small flakes.



Very complex crystal lattices

Quartz

Pink feldspar

Quartz is the simplest silicate mineral in chemical composition, consisting of nothing but interlocking silica tetrahedra. Because each oxygen is shared by two adjacent tetrahedra, the mineral as a whole has only two times rather than four times as many oxygens as silicones (the ratio in a single tetrahedron). Hence the chemical formula of quartz is  $\text{SiO}_2$ . Quartz is very hard because its silicones and oxygens are tightly bonded. Feldspars differ from quartz in that their structure includes both silica tetrahedra and tetrahedra in which aluminum takes the place of silicon. Ions of one or more additional types (potassium, sodium, and calcium) also fit into the framework in varying proportions. Unlike quartz, feldspars display good cleavage.

**FIGURE 2-9** Rock-forming silicate minerals. The diagrams show silica tetrahedra ( $\text{SiO}_4$ ) and their arrangements in three rock-forming silicate mineral groups. (Other atoms are omitted for

simplicity.) (Amphibole, Breck P. Kent; Mica, Quartz, and Pink feldspar, © 2013 National Museum of Natural History, Smithsonian Institution; Clay, Philippe Psaila/Science Photo Library/Science Source.)

particular body of rock originated. Sometimes an individual mineral tells the story, and sometimes it is the entire suite of minerals within the rock. Minerals within rocks found at the top of a mountain, for example, may reveal that those rocks originated deep below Earth's surface at high temperatures and pressures; such rocks have obviously been uplifted to their present elevation by the forces that formed the mountain. As another example, the oldest sedimentary rocks containing abundant iron minerals that are rich in oxygen have been taken to indicate when oxygen first built up to relatively high levels in Earth's atmosphere. Many of these rocks are between 2.4 billion and 1.8 billion years old, but some are much older.

### A few families of minerals form most rocks

The **silicates** are minerals containing *silica*, which is silicon bonded to oxygen. They are the most abundant group of minerals in Earth's crust and mantle. In silicates, four negatively charged oxygen atoms form a tetrahedral structure around a smaller, positively charged silicon atom. Figure 2-9 illustrates how silica tetrahedra unite in various ways, usually with other atoms, to form the most prominent silicate minerals of Earth's crust. Most silicate minerals form at high temperatures. Quartz, however, is also a common cement in sedimentary rocks and can form from warm or hot groundwater solutions. Silicates, including feldspar and quartz, are the primary constituents of the igneous and metamorphic rocks of Earth's crust. Because many silicate grains survive the weathering of these rocks and accumulate as sediments, most detrital sedimentary rocks also consist primarily of silicate minerals.

Carbonate and sulfate minerals also play large roles in the formation of rocks, but unlike silicates, most of these minerals form at low temperatures near Earth's surface. **Carbonate minerals** are constructed of one or more positive ions, such as calcium, magnesium, or iron, bonded to the complex ion  $\text{CO}_3^{2-}$ . The carbonate minerals calcite and aragonite are two forms of  $\text{CaCO}_3$  with different crystal structures (see Figure 2-7). **Dolomite** resembles calcite, but half of the calcium ions are replaced by magnesium, and its crystal structure has a special ordering in which calcium and magnesium ions are segregated into their own layers. Unlike calcite and aragonite, dolomite is not secreted by any organism in the form of a skeleton.

**Sulfate minerals** are formed of positive ions (such as calcium, barium, or strontium) that are attached to the complex ion  $\text{SO}_4^{2-}$ . As we will see later in this chapter, many sulfates are formed at low temperatures near Earth's surface through the evaporation of ocean or lake

water, during chemical weathering, and by precipitation from salty groundwater.

Although **oxides** make up only a small percentage of the large bodies of rock on Earth, these minerals form many important ore deposits. Rocks whose primary components are the oxides magnetite ( $\text{Fe}_3\text{O}_4$ ) and hematite ( $\text{Fe}_2\text{O}_3$ ), for example, yield most of the iron that is put to human use. Similar minerals, such as goethite ( $\text{FeO}(\text{OH})$ ), are common in soils and other low-temperature environments.

## Types of Rocks

Rocks are classified on the basis of their composition and the size and arrangement of their constituent grains. As we saw in Chapter 1, a body of rock belonging to any of the three basic types of rock—igneous, sedimentary, and metamorphic—can be transformed into another body of rock of the same type or of either of the other two types (see Figures 1-7 and 1-9).

### Igneous rocks form when molten rock cools

Igneous rocks are classified according to their chemical composition and grain size, both of which reflect a rock's mode of origin.

**Composition and density** Recall from Chapter 1 that most igneous rocks in continental crust fall into one of two major groups: felsic or mafic. Felsic rocks, which are rich in silica and aluminum, are generally light-colored and of low density. For this reason, continental crust is predominantly felsic. **Granite** is the most abundant kind of rock in Earth's continents (Figure 2-10). Two kinds of feldspar constitute about 60 percent of its volume; one of them gives many bodies of granite a pink color (see Figures 1-2 and 2-9). Felsic rocks also contain large percentages of quartz. Quartz is the silicate mineral with the highest concentration of silicon because, as  $\text{SiO}_2$ , it consists entirely of silica (silicon and oxygen; see Figure 2-9). Feldspar is also relatively rich in silica, although it also contains aluminum and either potassium, sodium, calcium, or a mixture of sodium and calcium.

Mafic igneous rocks, in contrast, are relatively low in silica and contain no quartz. Because mafic rocks, such as **gabbro** and **basalt** (see Figure 2-10) contain an abundance of magnesium, iron, or both, they are darker than felsic rocks. Iron also makes them denser. Basalt forms most of the oceanic crust, while ultramafic rocks, which are even lower in silica, form the mantle below the crust. Polished slabs of gabbro are sometimes sold commercially under the misleading name "black granite."

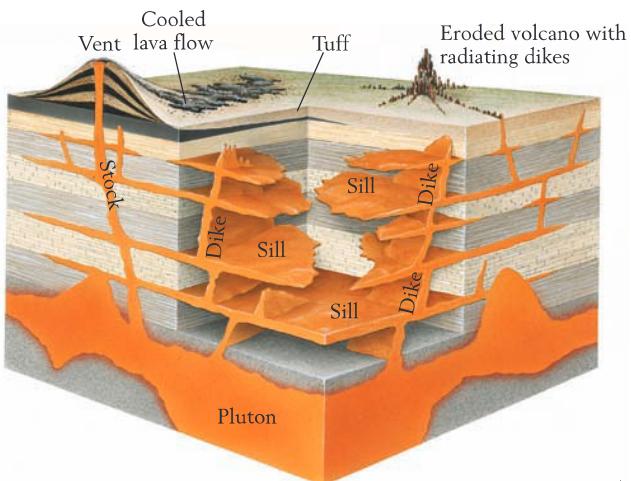


**FIGURE 2-10 Common igneous rocks.** Basalt (upper left) and rhyolite (lower left) are mafic and felsic extrusive rocks, respectively; they are fine-grained because they cooled rapidly at Earth's surface. Gabbro (upper right) and granite (lower right) are the intrusive equivalents, which are coarser-grained because they cooled slowly at great depth. (© 2013 National Museum of Natural History, Smithsonian Institution.)

**Cooling rate and grain size** Igneous rocks are also classified according to their texture, which is especially useful because their grain size reflects the rate at which they cooled from a molten state. Molten rock found within Earth is known as **magma**. If magma cools slowly, deep within the crust, its crystals can grow large, thus producing a coarse-grained rock. In contrast, rapid cooling, which takes place at and sometimes near Earth's surface, freezes molten rock into small crystals that yield a fine-grained rock (see Figure 2-10). Extremely rapid cooling produces glassy volcanic rock known as **obsidian**, one of the kinds of rock that Native Americans and other groups used to form arrowheads and cutting tools.

Most molten rock that cools within Earth's crust or at its surface comes from the mantle; as this molten rock rises, in the form of a blob or plume, it sometimes melts crustal rock with which it comes in contact. Magma rises through the crust in one of two ways. It can rise as a **diapir** by pushing aside overlying rock, or it can rise through a process known as **stoping**, whereby pieces of country rock (i.e., xenoliths) above the magma body break off and settle through the magma, thereby allowing it to move upward. The coarse-grained bodies of rock that magma forms when it cools within Earth are referred to as **intrusions** because they often displace or melt their way into preexisting rocks (Figure 2-11). They are also called **plutons**. **Sills** are sheetlike or tabular plutons that have been injected between sedimentary layers, and **dikes** are similarly shaped plutons that cut upward through sedimentary layers or crystalline rocks (see Figure 2-11A).

Molten rock that appears at Earth's surface through an opening, or **vent**, is called **lava**. Lava that has cooled to



A



B



C

**FIGURE 2-11 Configurations of extrusive and intrusive bodies of igneous rock on a continent.** A. Magma in large chambers can cool to form massive plutons. Magma that rises through a steeply sloping crack cools to form a dike. Magma that is injected between sedimentary strata cools to form a sill. B. Giant xenoliths of sandstone incorporated into the top of a mafic sill in the Dry Valleys of Antarctica. The largest xenolith is about 30 meters (100 feet) wide. C. Xenoliths of gabbro (dark) incorporated into a granite (light) pluton as part of the Keweenawan Mellen Intrusive complex (Midcontinent Rift System). Hammer for scale. (B, Ryan Currier, University of Wisconsin-Green Bay; C, John Luczaj, University of Wisconsin-Green Bay.)



**FIGURE 2-12** Recently cooled lava in Hawaii. The surface of the rock exhibits a ropy structure that reflects the pattern of the lava flow. (Peter L. Kresan.)

form solid rock often exhibits “frozen” flow structures on its surface (Figure 2-12). Some lavas erupt from tube-shaped vents to build cone-shaped volcanoes (see Figure 2-11A). One such structure is Mount St. Helens, which erupted in Washington State in 1980. A hollow crater forms at the summit of most volcanoes after an eruption as the unerupted lava sinks back down into the vent and hardens.

Other volcanic rocks form simply by flowing out of cracks, or **fissures**, from which they spread over large areas. These rocks are almost always mafic, because felsic lavas are more viscous (thicker liquids) and do not flow as easily. Mafic extrusive rocks that have flowed widely are often referred to as **flood basalts**. A flood basalt forms the broad Columbia Plateau in the northwestern United States (Figure 2-13).

Lava that emerges from the crust beneath the sea cools rapidly in a way that gives its surface a hummocky configuration, creating rock known as **pillow basalt**.



**FIGURE 2-13** A flood basalt. Flows of Miocene basalt formed this cliff along the Columbia River near Vantage, Washington. The cliff is about 330 meters (1100 feet) high. (Calvin Larsen/Science Source.)

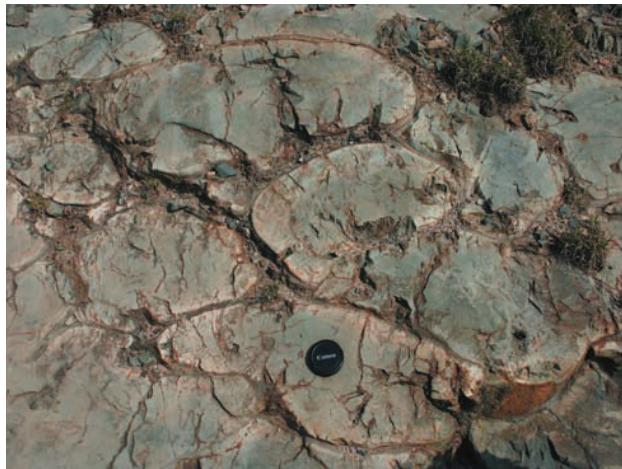
(Figure 2-14). Thus pillow structures in ancient basalt indicate that it cooled under water rather than on land.

Volcanic rocks can also form in ways other than by the cooling of flowing lava. Some volcanic eruptions—including that of Mount St. Helens in 1980—are explosive, hurling solid fragments of previously formed volcanic rock great distances. These fragments range from dust size to several meters across. Loose fragments of various sizes settle to form rock known as **tuff** (see Figure 2-11A). Although deposited in the same manner as sedimentary rock, tuff is usually classified as volcanic simply because it consists of volcanic particles. In fact, some tuffs form from hot grains that melt together as they settle, but others harden only after water percolating through them precipitates cement. Volcanoes sometimes eject frothy masses of lava that cool quickly to form the glassy rock known as **pumice**. Most pumice is so full of small gas pockets that it floats on water.



A

**FIGURE 2-14** Pillow basalt. A. Young pillow basalt preserved in Oman. B. This ancient greenstone (hydrothermally altered basalt) from Michigan still preserves the pillow structures that formed when it was erupted as basalt. The dark rims around the



B

greenish pillows come from alteration by seawater along the outside margin of the pillows. (A, Jessica M. Warren, Stanford University; B, John Luczaj, University of Wisconsin-Green Bay.)

Rocky material is not all that issues from volcanic vents. Gases of many kinds are also emitted, and in areas such as Yellowstone National Park, steamy geysers shoot up from sites where underground water is heated by magma and partially vaporized.

### Sedimentary rocks form from particles that settle through water or air

Most sedimentary rocks are formed of particles that belong to one of three categories: (1) fragments produced by the weathering and erosion of other rocks, (2) crystals precipitated from seawater, or (3) skeletal debris from organisms.

**Weathering, erosion, and sediment production** As we have seen, rocks at or near Earth's surface can be broken down by the physical and chemical processes that constitute weathering. Ice, snow, water, temperature changes, and earth movements are the primary agents of physical weathering. Water expands when it freezes, and when it freezes in cracks and crevices within rocks, it exerts such tremendous pressure that it can split the rocks apart. Repeated heating and cooling causes rocks to break apart, often along the boundaries of mineral grains. The products of weathering ultimately move away from the site of their origin under the influence of gravity, wind, water, or ice: the agents of erosion.

Destructive chemical processes constitute the most pervasive kind of weathering, and water and watery solutions act as their primary agents. Water at Earth's surface readily converts feldspar into clay, for example, carrying away some ions in the process. This process is often driven by naturally occurring carbonic acid, a weak acid produced when  $\text{CO}_2$  dissolves in rainwater. Like micas, clay minerals are sheet silicates (see Figure 2-9). Clay differs from mica, however, in that the molecular structure of its sheets is weaker. Thus clay minerals form only very small flakes, instead of large sheets like those that typify mica; that is why clay sediments have fine-grained textures. Carbonic acid is also the main agent responsible for cave formation and other *karst* processes, in which limestone or dolostone rocks are dissolved by slightly acidic waters at or near Earth's surface.

Quartz, in contrast to feldspar, is quite resistant to weathering. This characteristic, and the resistance of quartz to abrasion, accounts for the abundance of sand on Earth's surface. Most sand consists of quartz grains that are the same size as, or slightly smaller than, the quartz grains of granites and other crystalline rocks, from which most of them are derived.

When the feldspar grains of a crystalline rock such as granite weather to clay, the rock crumbles, releasing both flakes of clay and grains of quartz as sedimentary particles. Rainfall or meltwater washes many of these particles into streams, from which they are carried to larger bodies

of water. Eventually many of the particles settle from the waters of rivers, lakes, or oceans as sediment, which in time may become hard sedimentary rock.

Water, carbonic acid, and oxygen all take part in the weathering of mafic rocks, converting their iron-rich minerals into clay and iron oxide minerals that resemble rust. In the process, silica is carried away in solution. Mafic minerals are generally less stable at Earth's surface than felsic minerals and therefore undergo more rapid chemical weathering. Because most mafic minerals weather soon after exposure to air and water, these minerals are not abundant in most beach sands that fringe oceans. Feldspars, too, are rarely found on sandy beaches because most turn to clay either within or close to their parent rock.

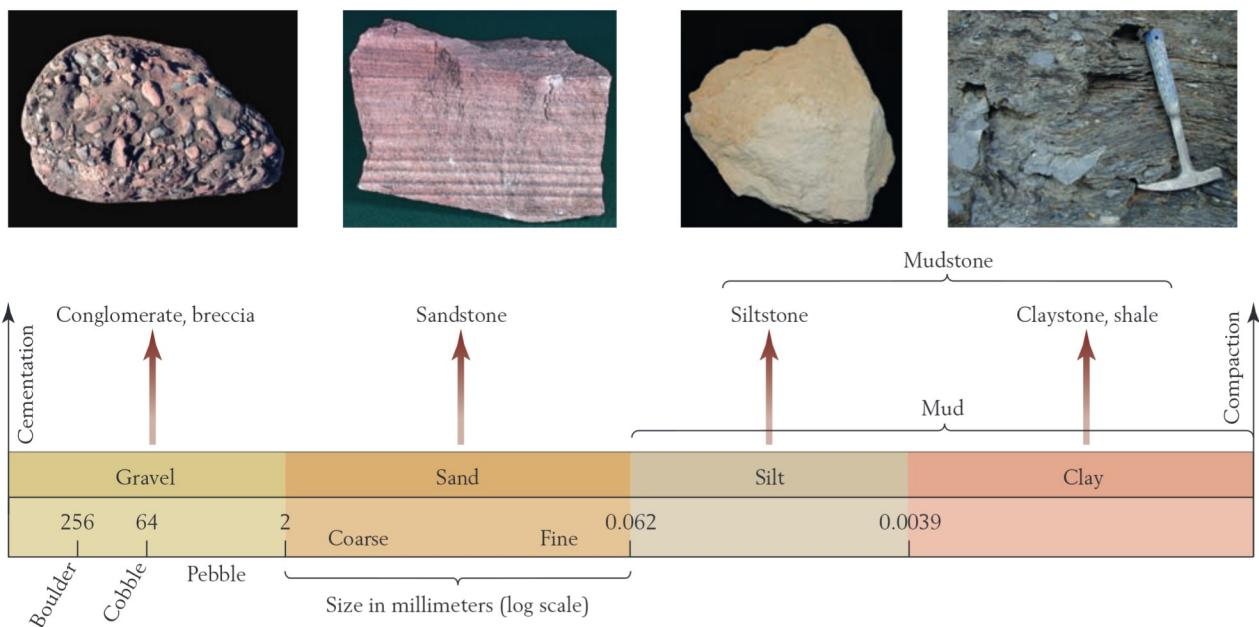
**Siliciclastic sedimentary rocks** Fragments of rock produced by destructive processes are termed **clasts**. **Siliciclastic** rocks, then, are sedimentary rocks composed of clasts of silicate minerals. Such rocks are very abundant in Earth's crust.

Siliciclastic sedimentary particles and rocks are classified according to grain size (Figure 2-15). The term **clay** is applied to particles smaller than 0.0039 millimeter. Although this quantitative definition may seem redundant and contradictory, because clays also constitute a family of sheet silicate minerals, the use of the term *clay* in reference to both mineralogy and particle size seldom creates confusion: nearly all clay mineral particles are of clay size, and nearly all particles of clay size are clay minerals. **Silt** is the name given to particles in the next size category, with diameters between 0.062 (or 1/16) and 0.0039 (or 1/256) millimeter. **Mud** is a term that embraces aggregates of clay and silt. Rocks formed largely of mud are termed **mudstones**. Mudstones that consist largely of clay tend to be **fissile**, meaning that they break along bedding surfaces; they are called **shale**. Their fissility results from the tendency of flakes to align horizontally during their deposition (Figure 2-16). Mudstones that contain an abundance of silt are less fissile than shale.

Particles of **sand** size range from 0.062 millimeter to 2 millimeters in diameter. When sand is cemented, it becomes sandstone (see p. 7). Most sedimentary grains of silt and sand size are composed of quartz. Even though quartz is very hard, quartz grains suffer some abrasion as they bounce and slide downstream along the floors of rivers. In the process, they tend to become smaller and more rounded.

**Gravel** refers to all particles larger than sand, including **pebbles**, **cobbles**, and **boulders**. A rock containing large amounts of gravel is termed a **conglomerate** if the gravel is rounded and a **breccia** if it is angular. In both cases, sand nearly always fills the spaces between the larger pieces of gravel.

When sediment settles from water, coarse-grained particles settle faster than fine-grained particles, as can be seen when sediments of various grain sizes are mixed

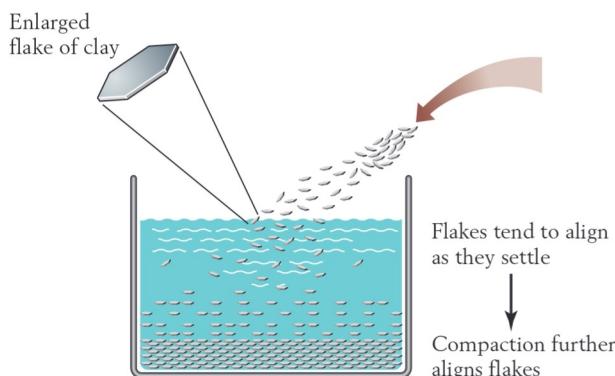


**FIGURE 2-15** Classification of sedimentary rocks according to grain size. Sediments range in size (given in millimeters) from clay to silt, sand, and gravel; gravel is divided into pebbles, cobbles, and boulders. Gravelly rocks are called conglomerates when their pebbles and cobbles are rounded and breccias when they are angular; these rocks normally contain sand as well. Rocks in which sand dominates are called sandstones. Rocks

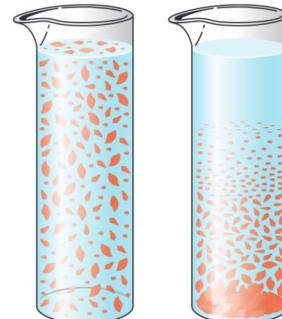
in which silt dominates are called siltstones. Rocks formed of clay are called claystone if they are massive and shale if they are fissile (or platy). Siltstones, claystones, and shales are all varieties of mudstone. All of the specimens shown here would fit in the palm of your hand. (Conglomerate, siltstone, and shale, John Luczaj, University of Wisconsin–Green Bay; Sandstone, Breck P. Kent.)

with water in a tall glass container and allowed to settle (Figure 2-17). Clay settles so slowly in water that very little of it falls from rapidly moving water, such as that of a stream or wave-ridden shallow sea. Most of it is deposited in calm waters, such as those of lakes, quiet lagoons, and the deep sea. Sand, in contrast, tends to accumulate along beaches and on the bottoms of swiftly flowing streams.

Coarse sediments not only settle more quickly from water than do fine sediments, but they are also less easily picked up or rolled along surfaces by moving water. Gravel, for example, tends to remain near its source area (the area where it was originally produced by erosion); thus gravel usually accumulates along the flanks of mountains from which it has eroded, seldom reaching the deep sea.



**FIGURE 2-16** Alignment of clay particles. Flakes of clay tend to assume a horizontal orientation as they settle through a body of water and pile up on the bottom. Those that initially lie at an angle tend to be aligned by compaction of the sediment.



**FIGURE 2-17** The settling pattern of sediment after it is suspended in water. The coarsest sediment settles most quickly and therefore ends up at the bottom of the deposit. The finest sediment settles last.

Different degrees of sediment sorting can be seen in Figure 2-18. Grains are said to be *poorly sorted* when they are of mixed sizes; the implication is that moving water did not separate the grains well according to size before they were deposited. Sand along a beach, in contrast, has usually been washed and transported by water currents and waves, and thus it tends to be *well sorted*—that is, its particles tend to fall within a narrow size range. Most of the particles in a handful of beach sand are likely to be either medium- or fine-grained (see Figure 2-15).

Siliciclastic rocks are classified according to composition as well as grain size. Because so many sand-sized grains are quartz, the word **sandstone** is sometimes automatically interpreted to mean quartz sandstone, but several other kinds of siliciclastic rocks also consist largely of sand-sized grains. One of them, **arkose**, contains at least 25 percent feldspar, which often gives it a pinkish color. Because feldspar tends to weather rapidly to clay, arkose usually accumulates only in proximity to its parent rock, soon after the feldspar grains are released by partial weathering and erosion. **Lithic sandstone** contains sand-sized grains that are made of rock fragments.



A



B

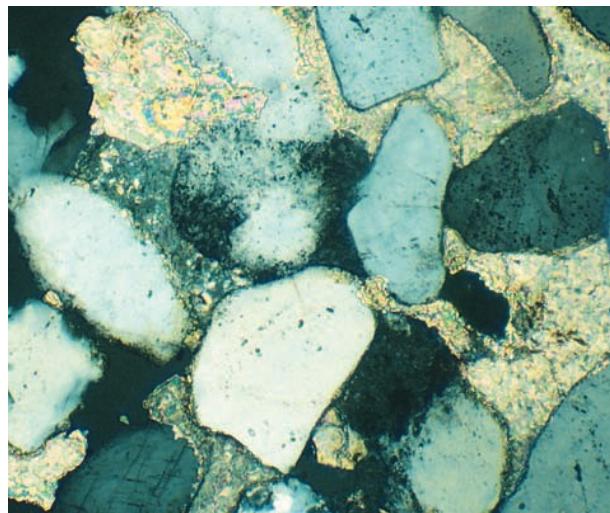
**FIGURE 2-18 Sediment sorting.** The photos show (A) poorly sorted and (B) well-sorted sand grains. (A, Rory Buckland RF/Alamy; B, AfriPics.com/Alamy.)

A black sand beach in Hawaii, for example, would make a lithic sandstone if its tiny fragments of basalt were cemented together. Lithic sands are common in volcanically active regions.

Another rock in which sand-sized particles predominate is **graywacke**, so called because it is usually dark gray. Graywacke consists of a variety of sedimentary particles, including sand- and silt-sized grains of feldspar and dark rock fragments as well as substantial amounts of clay. Most of the clay in graywacke was not carried to the environment of deposition in its present state, but was formed by the disintegration of larger grains within the rock.

Thus far we have discussed the nature of siliciclastic sediments, but not the ways in which these loose sediments become hard rock. A variety of lithification processes transform soft sediments into rock. The primary physical process of lithification is **compaction**, a process in which grains of sediment are squeezed together beneath the weight of overlying sediment. Muddy sediments usually undergo a great deal of compaction after burial as water is squeezed from them.

The most important chemical process of lithification is **cementation**. In this process, minerals precipitate from watery solutions that percolate through the pores between grains of sediment. The cement thus produced may or may not have the same chemical composition as the sediment. Sandstone, for example, is often cemented by quartz but more commonly by calcite. Cement can be observed microscopically in slices of rock ground thin enough to transmit light (Figure 2-19). Cementation is less extensive in clayey sediments than in clean sands because after clays undergo compaction, they are relatively



**FIGURE 2-19 Cement bordering quartz grains in sandstone, as seen in a thin section through a microscope.** Here the rounded sand grains appear gray or white in polarized light, depending on their orientation, and the crystalline calcite cement between the grains is iridescent yellow. (Peter L. Kresan.)



A

impermeable to mineral-bearing solutions. Clean sands, in contrast, are initially so permeable that their porosity is sometimes eliminated by pore-filling cement.

Cement sometimes gives sedimentary rocks their color. This is often the case for red siliciclastic sedimentary rocks called **red beds**. Some red beds are reddish because of an abundance of pink feldspar, but most derive their color from iron oxide, which acts as a cement.

**Chemical and biogenic sedimentary rocks** The grains of some sedimentary rocks are products of precipitation.

**Chemical sediments** result from precipitation of inorganic material from natural waters, sometimes as a result of evaporation. **Biogenic sediments** consist of mineral grains that were once parts of organisms. Some of these grains are pieces of skeletons, such as snail shells or colonies of coral, and others are the tiny, complete skeletons of single-celled creatures. Most biogenic sediments, however, consist of the skeletal remains of a variety of organisms, rather than just one or two. Because it is not always possible to determine whether a sedimentary rock is of chemical or biogenic origin, we will consider both groups together.

The most common chemical sedimentary rocks are **evaporites**, which form from the evaporation of seawater or other natural water. Many evaporites are extensive, well-bedded deposits that consist of vast numbers of crystals. Among the most abundant evaporites are **anhydrite** (calcium sulfate,  $\text{CaSO}_4$ ) and **gypsum** (calcium sulfate with two water molecules attached,  $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$ ). The terms *anhydrite* and *gypsum* refer both to the minerals with these names and to rocks that are composed largely of these minerals. **Halite** is another term that refers both to a mineral and to an evaporite rock. Recall that in arid regions with high rates



B

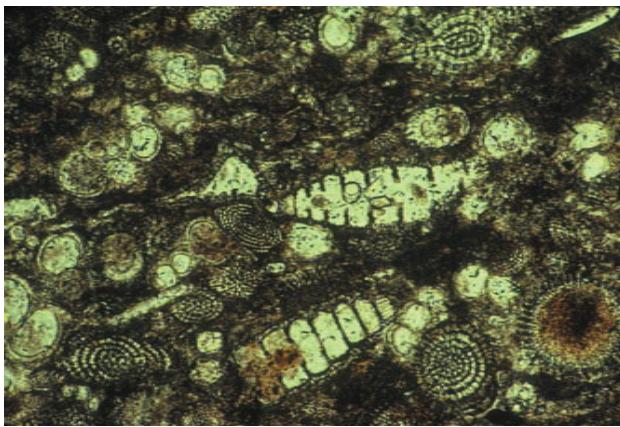
**FIGURE 2-20 Chert, a sedimentary rock composed of very small quartz crystals.**

A. Some chert is a chemical sedimentary rock and some is biogenic. B. Native Americans made arrowheads from chert, which is known informally as flint. (Breck P. Kent.)

of evaporation, ions released from rocks by weathering can become concentrated in lakes (see p. 7). Salty lakes in Death Valley, California, for example, lie atop halite-bearing deposits several kilometers thick that have accumulated over millions of years.

Just as evaporites are readily precipitated from water, they are readily dissolved, so they do not survive long at Earth's surface except in arid climates. Evaporites can survive for long geologic intervals, however, when they are buried deeply enough beneath younger deposits to be protected from fresh groundwater.

Other types of chemical sediments are less abundant than evaporites. Among the most common of these are chert, phosphate rocks, and iron formations. **Chert**, also called **flint**, is composed of extremely small quartz crystals that have precipitated from watery solutions. Typically, impurities give chert a gray, brown, red, or black color. Chert breaks along curved, shell-like surfaces (Figure 2-20); Native Americans took advantage of this feature when they fashioned chert into arrowheads. Some cherts occur as scattered, irregular masses in other kinds of sedimentary rocks. These masses, as well as some extensive beds of chert, grew from silica-rich solutions that moved through rock. Other bedded cherts are thought to have formed by direct precipitation of silica ( $\text{SiO}_2$ ) from seawater. Still others are biogenic deposits that result from the accumulation of the microscopic skeletons of single-celled organisms on the seafloor. These skeletons consist of a type of silica that differs from quartz in that it is amorphous (or noncrystalline). Over long geologic intervals, water percolates through deposits of these skeletons, converting them into very hard chert (Figure 2-21). Cherts older than 100 million years or so have suffered such extensive chemical alteration that many are difficult to identify as biogenic or chemical.



**FIGURE 2-21** Photomicrograph of Cretaceous age radiolarian-rich sediment from a Deep Sea Drilling Program core in the Pacific Ocean. Siliceous single-celled organisms known as radiolarians make up the majority of fossils in this thinly sliced rock, along with a few calcareous foraminifera (smaller circular fossils). The large elongated radiolarians are about 300 micrometers long. The dark brown material that makes up a small portion of the sample is clay-rich carbonate. Sediments made entirely of radiolarians often rapidly recrystallize to chert, obscuring many fossils. (Daniel Bernoulli, Universität Basel.)

**Banded iron formations** are complex rocks that consist of oxides, sulfides, or carbonates of iron interlayered with thin beds of chert. Banded iron formations are widespread only in very old Precambrian rocks, and some form large iron ore deposits.

**Limestones** include both chemical and biogenic bodies of rock. Because limestones are not as soluble in water as evaporites, they are much more common at Earth's surface, where they are quarried extensively for the production of building stone, gravel, and concrete. We have already noted that although ancient limestones consist primarily of the mineral calcite, many of their grains were initially composed of aragonite that was transformed into calcite over time (see p. 31).

Dolomite is a carbonate mineral that is relatively uncommon in modern marine environments, but is common in many ancient rocks. As we have seen, it differs from calcite in that half of the calcium ions of its crystal structure are replaced by magnesium ions. In fact, much dolomite has been formed by the chemical alteration of calcite. When dolomite is the dominant mineral of rock, the rock, too, is sometimes called dolomite, but is more properly labeled **dolostone**. Because limestones and dolostones are similar rocks and are often found closely associated, they are sometimes referred to collectively as **carbonate rocks**. Similarly, unconsolidated sediments consisting of aragonite, calcite, or both are often called **carbonate sediments**.

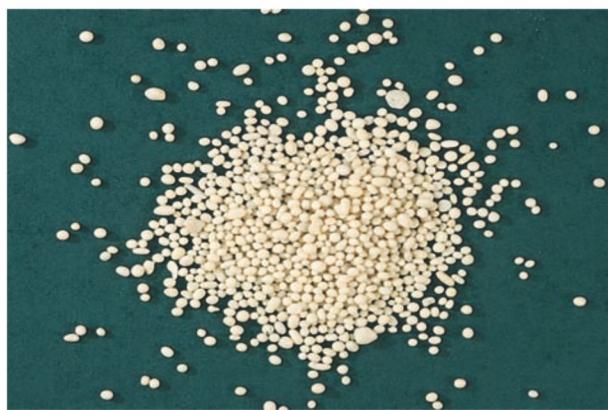
Carbonate sediments form in two ways: by direct precipitation from seawater and through accumulation of skeletal debris from organisms. Many types of marine life grow shells or other kinds of skeletons that consist of aragonite or calcite. These organisms contribute skeletal material to the seafloor as sedimentary particles. Some of

these particles retain their original sizes, while others become smaller through breakage or wear. The product of this biological contribution is an array of clastic particles that are similar to siliciclastic grains and, like them, can be classified according to size. Thus we speak of carbonate sands and carbonate muds.

Most discrete particles in limestone that are sand-sized or larger can be seen to be skeletal particles (Figure 2-22). The origin of mud-sized material is more difficult to determine. The main constituents of carbonate muds in modern seas are **aragonite needles** (see Figure 2-7B), which are produced both by direct precipitation and by the disintegration of carbonate skeletons, especially of algae. In ancient fine-grained limestones, aragonite needles have been transformed into tiny calcite grains. The resulting granular texture reveals little about the configuration or mode of origin of the original carbonate particles.



**FIGURE 2-22** Cross-section of a fossiliferous limestone. This rock consists largely of skeletal debris from organisms known as crinoids. These disc-shaped fragments are portions of the animal's "stem" that secured it to the seafloor, but broke apart after the organism's death. Carbonate mud and calcite cement make up the dark material between the framework grains. The largest grains here are about 1 centimeter (0.4 inches) wide. (David Lyons/Alamy)



A

**FIGURE 2-23 Ooids form oolite rocks.** A. These nearly spherical ooids, which are the size of sand grains, have formed on the modern seafloor. B. A photomicrograph of thinly sliced oolite



B

rock reveals the concentric structure of the ooids. (A, Steven M. Stanley; B, Courtesy Mark A. Wilson, Department of Geology, The College of Wooster.)

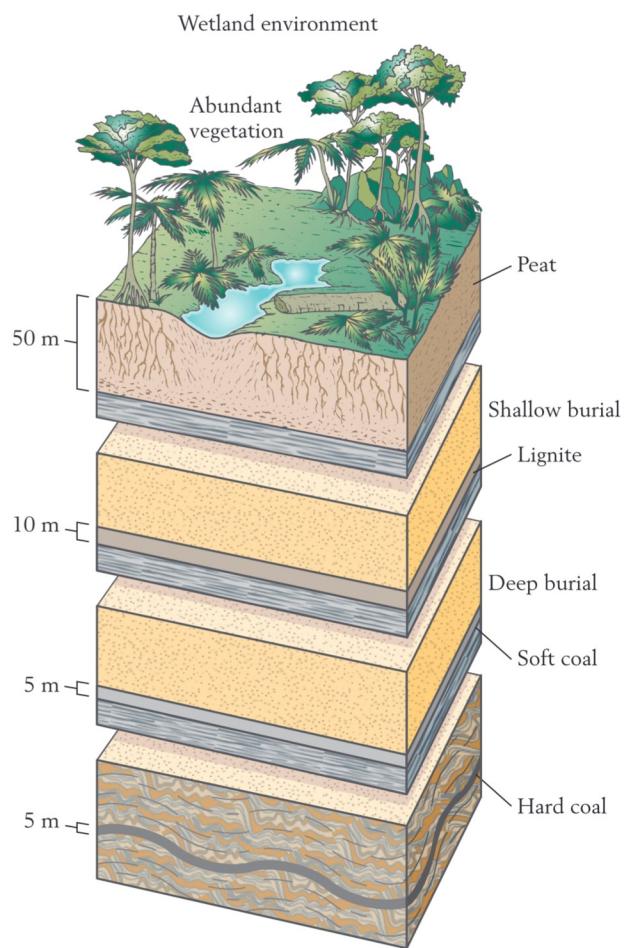
**Oolites** are sediments or rocks consisting of nearly spherical grains (*ooids*) that can be seen to grow in modern seas by rolling about and accumulating aragonite needles, in the same way a snowball rolled about in the snow becomes large enough to make a snowman (Figure 2-23A). Ooids form in shallow water where the seafloor is agitated by strong water movements. A cross section of an ooid displays a concentric structure, with thin, dark bands representing breaks in deposition when the grain was at rest and ceased to grow for a time (Figure 2-23B).

Other types of sand-sized (or larger) grains are less common in limestone, but can yield important information about ancient environments. **Intraclasts**, for example, are fragments of preexisting seafloor material that gets ripped up during storms or other sudden events. **Fecal pellets** are sand-sized grains found in carbonate mud that have passed through the guts of invertebrate organisms. These and other carbonate grain types yield particularly important information about life or physical conditions that existed in the area where they are found.

Calcium carbonate is precipitated from seawater only if the water contains relatively little carbon dioxide. Carbon dioxide combines with water to form carbonic acid, which causes calcium and carbonate ions to be retained in solution. Carbon dioxide is less soluble in warm than in cold water, so carbonate sediments accumulate primarily in tropical seas, where winter water temperatures seldom drop below 18°C (64°F). This phenomenon also accounts for the fact that few cold-water organisms secrete massive skeletons of calcium carbonate. Carbonate sediments can also form in freshwater habitats, usually as a result of the carbonate-secreting activities of certain algae.

Carbonate sediments can be compacted after burial, and they can also harden through cementation, in the same manner as siliciclastic sediments. Carbonate sediments are nearly always cemented by carbonate minerals simply because rich sources of such cements are close at hand.

**Coal** is rock formed of stratified plant debris (Figure 2-24). It can be burned because organic carbon compounds account for more than 50 percent of its



**FIGURE 2-24 The origin of coal.** Peat, which accumulates in water where little oxygen is present, is compressed and heated through burial. If buried deeply enough, it may eventually become hard coal.



**FIGURE 2-25** **A graded bed.** When sediments consist of grains of mixed sizes, the coarser grains settle more rapidly than the finer ones, so that grain size diminishes from bottom to top. (Steven M. Stanley.)

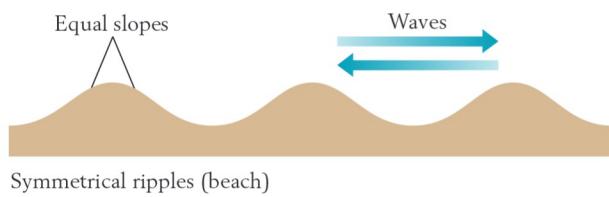
composition. The starting material for the formation of coal is **peat**, leafy and woody plant tissue that accumulates in oxygen-deprived water occupied by few bacteria that can cause decay. The heat and pressure of burial turn peat into brown coal called *lignite* through compression and expulsion of water, hydrogen, and nitrogen. The result is an increase in the carbon content of the organic material. Deeper burial continues the process, producing soft coal.

**Sedimentary structures** Distinctive arrangements of grains in sedimentary rocks are termed **sedimentary structures**. Sedimentary structures reflect modes of deposition and provide useful tools for interpreting the environments in which ancient sediments were deposited.

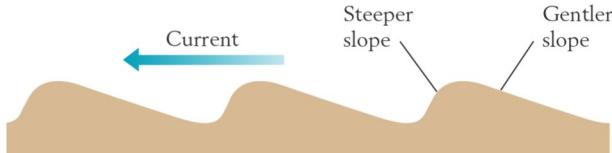
A **graded bed** is a sedimentary structure in which grain size decreases from the bottom to the top (Figure 2-25). This pattern usually results from the normal settling process that characterizes sediments of mixed grain size, in which the coarser sediment settles more rapidly than the finer sediment (see Figure 2-17). Most graded beds form in nature when a strong current suddenly introduces a large volume of sediment to a quieter body of water, where it settles.

**Ripples** formed on a beach have symmetrical cross sections because wave motion oscillates back and forth (see Figure 1-1). Water currents and wind, in contrast, produce ripples and larger ridges—bars or dunes—with asymmetrical cross sections. Sediment accumulates along the lee (downstream) side of such a ridge (Figure 2-26). The resulting accumulation of beds on that side produces what is termed **cross-bedding** or **cross-stratification** because sets of parallel beds slope at an angle to the horizontal. Although the sets of cross-bedding in a single rock unit vary somewhat in their orientation, their average direction of slope indicates the general direction of the prevailing winds when the ancient beds were deposited to form a dune, or of currents when the beds were deposited in a stream.

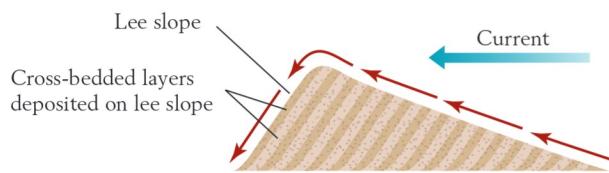
**Mudcracks** form as fine-grained, clay-rich sediments dry out and shrink. These structures, which often form patterns like the hexagonal tiles of a bathroom floor, are visible in ancient rocks even if sediment fills them. Mudcracks indicate deposition in shallow waters, such as those at the edge of a lagoon or lake, that receded and exposed the sediments to the air.



A



Asymmetrical ripples (dunes or river bars)



B



C

**FIGURE 2-26** **Ripples and cross-bedding.** A. Profiles of ripples produced by waves and by currents. B. The formation of cross-bedding on the lee slope of a ripple. C. Cross-bedding in sand formed, as in B, by water currents flowing toward the left. The

beds here are 2–3 centimeters thick. (A and B after F. Press, R. Siever, J. Grotzinger, and T. H. Jordan, *Understanding Earth*, 4th ed., W. H. Freeman and Company, New York, 2004; C, John Luczaj, University of Wisconsin–Green Bay.)

**Bioturbation** is the disturbance of sedimentary bedding by the activity of animals. Even a footprint disturbs the bedding directly beneath it, but animals that burrow in the seafloor have produced most of the bioturbated sediments observed in the rock record. These animals, most of which have been worms, have often pushed the distinctive sediment of a layer upward or downward, so that it has displaced an adjacent layer, or they have created hollow spaces into which sediment from the overlying layer has settled (Figure 2-27).

### Metamorphic rocks form from other rocks at high temperatures and pressures

Recall that metamorphic rocks form by the alteration of other rocks at temperatures and pressures that exceed those normally found at Earth's surface. Metamorphism alters rocks without melting them, whereas igneous rocks are products of the cooling of melted rock. Metamorphism alters both the mineral composition and the texture of all kinds of rocks—igneous, sedimentary, and metamorphic. The mineral assemblages of metamorphic rocks serve as critical “thermometers” and “barometers” because they vary with the temperature and pressure of metamorphism. **Grade** is the word used to indicate the levels of temperature and pressure of metamorphism. High-grade assemblages of metamorphic minerals form at higher temperatures and pressures than low-grade assemblages.

**Regional metamorphism** As its name implies, **regional metamorphism** transforms deeply buried rocks over distances of hundreds of kilometers. Igneous activity usually extends along the length of an actively forming mountain chain, supplying heat for metamorphism. Extending outward along each side of such an igneous belt is a zone of regional metamorphism produced by high temperatures and pressures. Most rocks in zones of regional metamorphism display a texture known as **foliation**, which is an alignment of platy minerals caused by the pressures applied during metamorphism. The grade of metamorphism in a regional metamorphic



**FIGURE 2-27** **Bioturbation in a limestone.** Remnants of bedding are visible, along with some finger-sized burrows, but disruption and mixing of what were originally alternating light and dark sediment layers have given much of the rock a mottled appearance. (John Luczaj, University of Wisconsin–Green Bay.)

zone typically declines with distance from the neighboring belt of igneous activity.

The following three types of rock represent different grades of metamorphism. **Slate** is a fine-grained rock of very low metamorphic grade in which foliation produces fissility much like that of shale (Figure 2-28A). Slate broken into rectangular plates covers the roofs of many



A



B



C

**FIGURE 2-28** **Foliated metamorphic rocks.** A. Slate is a low-grade metamorphic rock. B. Schist is a medium-grade metamorphic rock. C. Gneiss is a high-grade metamorphic rock.

(A, Breck P. Kent; B, John Luczaj, University of Wisconsin–Green Bay; C, John Luczaj, University of Wisconsin–Green Bay.)

houses. **Schist** is a low- to medium-grade metamorphic rock that consists largely of grains of platy minerals, often including mica (Figure 2-28B); because of its strong foliation, schist tends to break along parallel surfaces. **Chlorite** is a mica-like green mineral that occurs primarily in schist. When abundant, it gives the rock the informal name *greenschist*. **Gneiss** is a high-grade metamorphic rock whose intergrown crystals resemble those of igneous rock, being granular rather than platy, but whose minerals tend to be segregated into wavy layers (Figure 2-28C).

Not all rocks in regional metamorphic zones are foliated, however. Some have homogeneous granular textures, which indicate not only that their interlocking mosaics of crystals lack preferred orientations, but also that certain minerals are not segregated into bands. Marble and quartzite, for example, are usually homogeneous granular metamorphic rocks (Figure 2-29). **Marble**, which consists of calcite, dolomite, or a mixture of the two minerals, forms from the metamorphism of sedimentary carbonates. Marble is popular as a decorative stone, not only because impurities often create attractive patterns within it, but also because it is relatively soft and easy to cut and polish. **Quartzite**, which consists of nearly pure quartz, is much harder than marble; it forms from the metamorphism of quartz sandstone, during which quartz grains fuse together under high pressure.

**Contact metamorphism** When an igneous intrusion “bakes” the surrounding rock (see Figure 1-9), the result is **contact metamorphism**. High temperature plays a larger role in this process than high pressure. The resulting rocks are usually fine-grained. Contact metamorphism may occur deep within Earth or near the surface. Though a more localized phenomenon, it resembles

regional metamorphism in displaying a gradient: the grade of metamorphism declines away from the heat source.

**Hydrothermal metamorphism** The percolation of hot, watery fluids through rock can result in **hydrothermal metamorphism**. Fluids of this type escape when magma intrudes continental crust, but most hydrothermal metamorphism takes place along mid-ocean ridges, where seawater circulates through hot, newly formed lithosphere. Here basalt (see Figure 2-10) is extensively altered to greenstone (see Figure 2-14B). Many valuable ore minerals, including gold deposits, have been emplaced in other rock by hydrothermal fluids.

**Shock metamorphism** When an asteroid strikes Earth, rocks that border the impact crater are metamorphosed by the heat and very high pressure of the impact in what is termed **shock metamorphism**.

**Fault zone metamorphism** When rocks are ground up along a fault as a result of the intense shearing forces and heat of Earth movement, a distinctive linear zone of brecciated rock is formed. Recognition of such **fault zone metamorphism** helps geologists to reconstruct former plate boundaries and interpret the tectonic setting.

**Burial metamorphism** Rocks are altered when they are buried so deeply that they are exposed to temperatures and pressures high enough to change their mineralogical composition. For siliciclastic rocks, the consequences of this **burial metamorphism** are similar to those of regional metamorphism. Deep burial of soft coal can turn it into hard coal (see Figure 2-24). In the process, much water, hydrogen, and nitrogen are driven off. Hard coal is considered a metamorphic rock because the high temperatures and pressures that form it produce a new



A

**FIGURE 2-29** Nonfoliated metamorphic rocks. A. A typical marble specimen. Its width is about 13 centimeters (5 inches). B. A quartzite specimen from Australia. Its width is about 10



B

centimeters (4 inches). (A, Susan E. Degginger/Alamy; B, Joyce Photographics/Science Source.)

mineral as one of its major components. This mineral is graphite, which consists of pure carbon (see p. 32).

**The upper limit of metamorphism** When conditions become so hot that a rock melts, metamorphism ceases. Rocks that form when the resulting molten material cools are classified as igneous. Very high-grade metamorphic rocks that come close to a molten state before cooling resemble coarse-grained igneous rocks in being granular and lacking foliation. In fact, these metamorphic rocks are difficult to distinguish from igneous rocks.

## CHAPTER SUMMARY

### What traits of minerals determine their physical properties?

Minerals are elements or chemical compounds that are the building blocks of rocks. An isotope is a form of an element that differs in number of neutrons (and, hence, in atomic weight) from other isotopes of the same element. Some unstable isotopes are radioactive, which means that their nuclei decay to form other isotopes. A mineral's hardness is determined by the strength of its chemical bonds; its density is determined by the weight of its atoms and by their degree of packing within the crystal; and its pattern of fracturing reflects planes of weakness within its crystal structure. Most rocks of Earth's crust and mantle are formed by a few families of minerals, of which the most abundant is the silicate family, in which the basic building block is a silicon atom surrounded by four oxygen atoms.

### What conditions produce various kinds of igneous rock?

Mafic igneous rocks are denser than felsic igneous rocks, primarily because they are rich in minerals that contain iron, which has a high atomic weight. Intrusive igneous rocks are coarse-grained because they form from magma that cools slowly below Earth's surface. Extrusive igneous rocks are fine-grained because the molten lava from which they form cools and solidifies quickly at Earth's surface.

### What are the ways in which sedimentary rocks form?

Siliciclastic sedimentary rocks consist of silicate grains that are products of weathering and erosion. Weathering of granite, the most abundant kind of rock in continental crust, releases grains of quartz sand, the dominant constituent of sandstone, and clay, the dominant constituent of shale. Chemical sediments are formed by precipitation of inorganic material from natural waters; evaporites, for example, precipitate from water as it evaporates. Biogenic

sediments consist of mineral grains that were once parts of the skeletons of organisms. The most abundant biogenic sedimentary rock is limestone, which consists largely of calcium carbonate; limestone also forms by precipitation of carbonate from seawater. Some geologically young limestones consist of the chemically unstable mineral aragonite, but old limestones consist of the more stable mineral calcite. Distinctive arrangements of grains, known as sedimentary structures, reflect the mode of deposition of sedimentary rocks.

### How do metamorphic rocks form from other rocks?

Metamorphic rocks vary in composition and texture depending on the degrees of heat and pressure responsible for their origin. Regional metamorphism alters rocks over hundreds of kilometers, whereas contact metamorphism results when the heat of an igneous intrusion bakes local rock. Hydrothermal metamorphism results when hot, watery fluids percolate through rocks. Burial metamorphism of plant debris produces coal.

## REVIEW QUESTIONS

1. How does an ion form from an uncharged atom?
2. What allows one element to substitute for another in the crystal of a particular mineral?
3. How does the orientation of a dike of igneous rock differ from the orientation of a sill?
4. Under what circumstance does the eruption of lava produce a flood basalt instead of a volcano?
5. Why do most sediments formed by weathering consist of silicate minerals?
6. How does the origin of clay particles from igneous rock differ from the origin of grains of quartz sand from the same kinds of rock?
7. How does the process that turns clay into shale differ from the process that turns sand into sandstone?
8. Why do evaporite deposits weather quickly?
9. Explain how the following sedimentary structures form: (a) graded beds, (b) ripples, (c) cross-bedding, (d) mudcracks.
10. How does the origin of a metamorphic rock differ from the origin of an igneous rock?
11. Water plays a major role in the origin of sedimentary rocks. Use the Visual Overview on pages 26–27 as a guide to explore the ways in which water serves as a medium for the production, transport, deposition, and lithification of the various kinds of mineral grains that form sediment.