3   Minerals

We live in a world of minerals—they are everywhere around us. Gems and jewelry are minerals. Gravel and sand are minerals. Mud is a mixture of microscopic minerals. Ice is a mineral, and even dust in the air we breathe is made up of tiny mineral grains. Minerals sustain our lives and provide continuously for society. The houses in which we live, the automobiles we drive, as well as the roads and other structures of our society, and almost everything we touch are made of minerals or material derived from minerals. Indeed, on average, every person on Earth uses, directly or indirectly, 10 metric tons of minerals each year.

But the importance of minerals extends far beyond their value as economic deposits. Minerals are also the substance of Earth's natural systems. The green and white crystals in this beautiful photograph are two very different minerals. The lustrous pastel green crystals are apophyllite and the sparkling white needles are mesolite. Each mineral has distinguishing properties. Every one of the tiny ice-clear crystals in these radial sprays of mesolite has much in common with all of the other grains of its mineral species. For example, all grains of mesolite have the same internal arrangement of atoms and have the same chemical and physical properties even though individuals may vary greatly in size and shape. The atomic

structure of mesolite creates a natural chemical sieve. Its open structure allows some mole-cules and ions dissolved in water to move through the framework of the atoms, but it will filter out the larger molecules. Mesolite's internal structure contains chains of atomic tetra-hedrons aligned in one direction; this produces the needle-like shape of the crystals. The mineral breaks preferentially between the long chains where atomic bonds are weakest.

All of Earth's dynamic processes involve the growth and destruction of minerals as matter changes from one state to another. As Earth's surface weathers and erodes, some minerals are destroyed and others grow in their place. Mesolite and apophyllite in this photo grew from a watery solution as flowed through ancient lava flows. As sediments accumulate in the oceans, minerals also grow from solution. Other minerals grow from molten rock when lava erupts from volcanoes and cools. Deep below Earth's surface, high pressure and temper-ature remove atoms from the crystal structures of some minerals and cause them to recom-bine them into new minerals. As tectonic plates move and continents drift, minerals are creat-ed and destroyed by a variety of processes. Some knowledge of Earth's major minerals, therefore, is essential to understanding Earth's dynamics.

In this chapter, we survey the general characteristics of minerals and the physical prop-erties that identify them. We then explore the major rock-forming minerals in preparation for a study of the major rock types in Chapters 4, 5, and 6.

Photograph by Chip Clark.

1. An atom is the smallest unit of an element that possesses the properties of the element. It consists of a nucleus of protons and neutrons and a surrounding cloud of electrons.
2. An atom of a given element is distinguished by the number of protons in its nucleus. Isotopes are varieties of an element, distinguished by the different numbers of neutrons in their nuclei.
3. Ions are electrically charged atoms, produced by a gain or loss of electrons.
4. Matter exists in three states: (a) solid, (b) liquid, and (c) gas. The differences among the three are related to the degree of ordering of the atoms.
5. A mineral is a natural solid possessing a specific internal atomic structure and a chemical composition that varies only within certain limits. Each type of mineral is stable only under specific conditions of temperature and pressure.
6. Minerals grow when atoms are added to the crystal structure as matter changes from the gaseous or the liquid state to the solid state. Minerals dissolve or melt when atoms are removed from the crystal structure.
7. All specimens of a mineral have well-defined physical and chemical properties (such as crystal structure, cleavage or fracture, hardness, and density).
8. Silicate minerals are the most important minerals and form more than 95% of Earth's crust. The most important silicates are feldspars, micas, olivines, pyroxenes, amphiboles, quartz, and clay minerals. Important nonsilicate minerals are calcite, dolomite, gypsum, and halite.
9. Minerals grow and are broken down under specific conditions of temperature, pressure, and chemical composition. Consequently, minerals are a record of the changes that have occurred in Earth throughout its history.
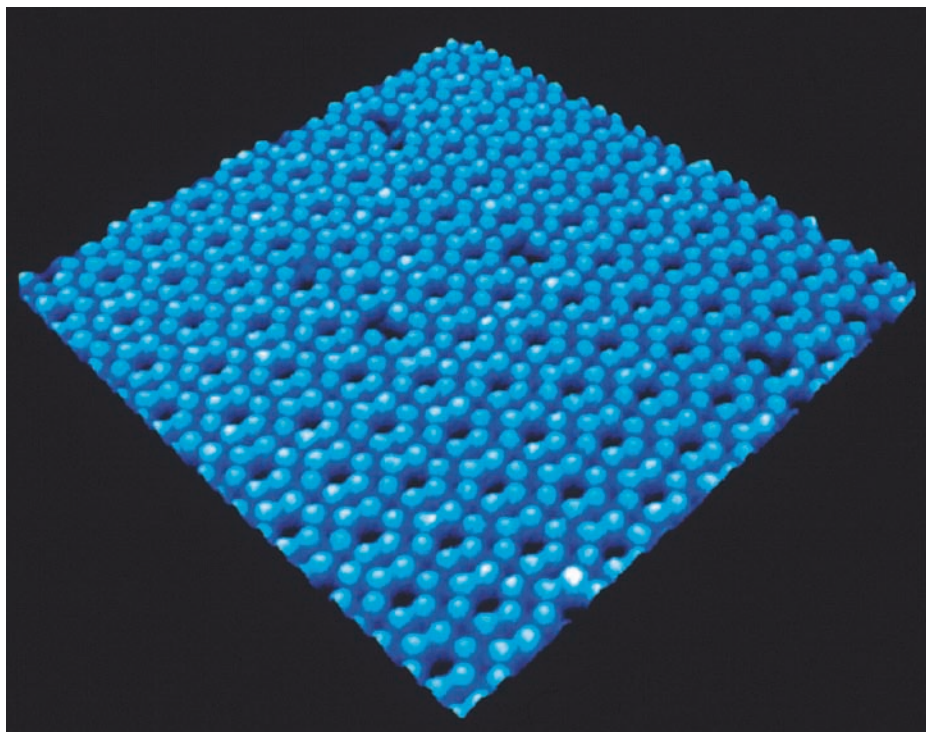
MATTER

An atom is the smallest unit of an element that possesses the properties of the element. It consists of a nucleus of protons and neutrons and a surrounding cloud of electrons. There are three states of matter: gas, liquid, and solid. Each state is distinguished by unique physical properties. Processes in Earth's dynamics mostly involve the changing of matter from one state to another.

To understand the dynamics of Earth and how rocks and minerals are formed and changed through time, you must have some knowledge of the fundamental structure of matter and how it behaves under various conditions. The solid materials that make up Earth's outer layers are called rocks. Most rock bodies are mixtures, or aggregates, of minerals. A mineral is a naturally occurring compound with a definite chemical formula and a specific internal structure. Because minerals, in turn, are composed of atoms, to understand minerals we must understand something about atoms and the ways in which they combine.

Atoms

An **atom** is the smallest fraction of an element that can exist and still show the characteristics of that element. Atoms are best described by abstract models constructed from mathematical formulas involving probabilities. They are much too small to be seen with optical microscopes; recently, however, images of atoms have been made. An example is shown in Figure 3.1. In its simplest form, an atom is characterized by a relatively small **nucleus** of tightly packed protons and neutrons, with a surrounding cloud of electrons. These are the principal building blocks of atoms, but many other subatomic particles have been identified in recent years.
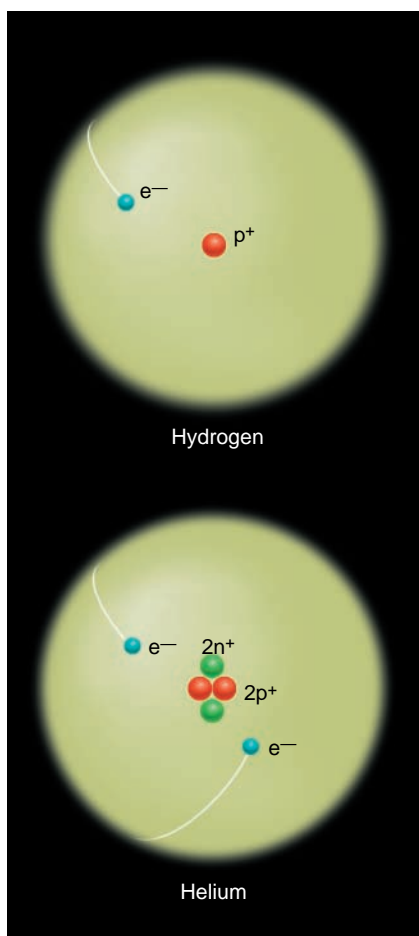
FIGURE 3.1   Image of atoms of silicon produced by a scanning tunneling microscope at the IBM Research Center, Yorktown Heights, New York. The blue spots are individual silicon atoms, which are arranged in a regular pattern that repeats itself across the surface. You can also see the hexagonal arrangement of groups of the atoms. Locally, flaws in the structure are also visible. Images such as this are helpful in understanding the structure of different minerals. *(Courtesy of International Business Machines Corporation. Unauthorized use not permitted.)*

The distinguishing feature of an atom of a given element is the number of protons in the nucleus. The number of electrons and neutrons in an atom of a given element can vary, but the number of protons is always the same. Each **proton** carries a positive electrical charge, and the mass of a proton is taken as the unit of atomic mass, approximately $1.66 \infty 10^{-24}$ g. The **neutron,** as its name indicates, is electrically neutral and has approximately the same mass as the proton. The **electron** is a much smaller particle, with a mass approximately 1/1850 the mass of the proton. It carries a negative electrical charge equal in intensity to the positive charge of the proton. Because the electron is so small, for practical purposes, the entire mass of the atom is considered to be concentrated in the protons and neutrons of the nucleus. The **atomic mass** is simply the sum of the number of neutrons and protons.

What is the structure of an atom?

Hydrogen is the simplest of all elements. It consists of one proton in the nucleus and one orbiting electron (Figure 3.2). The next heaviest atom is helium, with two protons, two neutrons, and two electrons. Each subsequently heavier element contains more protons, neutrons, and electrons. Figure 3.3 is a simplified chart of all naturally occurring elements. The elements are arranged in rows, with increasingly heavier elements to the right and bottom. This table is commonly called the periodic chart. The distinguishing feature of an element is the number of protons in the nucleus of each of its atoms, often called the **atomic number.** The number of electrons and neutrons in the atoms of a given element can vary, but the number of protons is constant.

Atoms normally have the same number of electrons as protons and thus do not carry an electrical charge. As the number of protons increases in progressively heavier atoms, the number of electrons also increases. The electrons fill a series of energy-level shells around the nucleus, each shell having a maximum capacity. The progressive filling of these shells is reflected in the rows of the periodic chart (Figure 3.3). The electrons in the outer shells control the chemical behavior of the element.

FIGURE 3.2   **The atomic structures of hydrogen and helium** illustrate the major particles of an atom. Hydrogen has one proton *(p)* in a central nucleus and one orbiting electron *(e)*. Helium has two protons *(p),* two neutrons *(n)* in the nucleus, and two orbiting electrons.

What are the distinguishing characteristics of an isotope? Of an ion?

Isotopes

Although the number of protons in each atom of a given element is constant, the number of neutrons in the nucleus can vary. This means that atoms of a given element are not all exactly alike. Iron atoms, for example, have 26 protons but individual atoms may have 28, 30, 31, or 32 neutrons. These varieties of iron are examples of **isotopes;** they all have the properties of iron but differ from one another in mass. Most common elements exist in nature as mixtures of isotopes. Some isotopes are unstable, emitting particles and energy as they experience radioactive decay to form new, more stable isotopes.

Ions

Atoms that have as many electrons as protons are electrically neutral, but atoms of most elements can gain or lose electrons in their outermost shells. If electrons are gained or lost, an atom loses its electrical neutrality and becomes charged. These electrically charged atoms are **ions.** The loss of an electron makes a positively charged ion because the number of protons then exceeds the number of negatively charged electrons. If an electron is gained, the ion has a negative charge. The electrical charges of ions are important because the attraction between positive ions and negative ions is the bonding force that sometimes holds matter together. Like atoms, ions have distinctive sizes that reflect the number of particles in the nucleus and the number of electrons in the surrounding cloud. Ionic size and ionic charge control how elements fit together to make solid minerals (Figure 3.3).
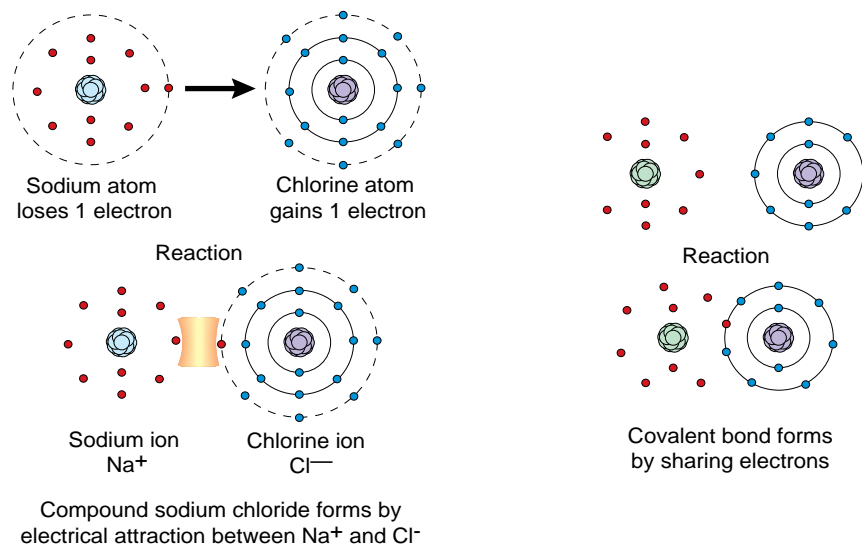
Bonding

An atom is most stable if its outermost shell is filled to capacity with electrons. The inner shell can hold no more than 2 electrons. The next shell can hold 8 electrons and is full in neon (atomic number 10). In heavier elements, the next shell can have 18 electrons, and the shell after that one can have 32 electrons. Neon, for example, has 10 protons in the nucleus and 10 electrons, of which 2 are in the first shell and 8 are in the second shell. A neon atom does not have an electrical charge. Its two electron shells are complete because the second shell has a limit of 8 electrons. As a result, neon does not interact chemically with other atoms. Argon and the other noble gases (the right column on the periodic chart) also have 8 electrons in their outermost shell, and they normally do not combine with other elements. Most elements, however, have an incomplete outermost shell. Their atoms readily lose or gain electrons to achieve a structure like that of argon, neon, and the other inert gases, with 8 electrons in the outermost shell.

For example, an atom of sodium has only 1 electron in its outermost shell but 8 in the shell beneath (Figure 3.4). If it could lose the lone outer electron, the sodium atom would have a stable configuration like that of the inert gas neon. The chlorine atom, in contrast, has 7 electrons in its outermost shell, and if it could gain an electron, it too would attain a stable electron configuration. Whenever possible, therefore, sodium gives up an electron and chlorine gains one. The sodium atom thus becomes a positively charged sodium ion, and the chlorine atom becomes a negatively charged chloride ion. With opposite electrical charges, the sodium ions and chloride ions attract each other and bond together to form the compound sodium chloride (common salt, also known as the mineral halite). (A **compound** has more than one element in its structure.) This type of bond, between ions of opposite electrical charge, is known as an **ionic bond.** Such bonds commonly develop between elements that lie far from one another on the periodic table.

Atoms can also attain the electron arrangement of a noble gas, and thus attain stability, by sharing electrons. No electrons are lost or gained, and no ions are formed. Instead, an electron cloud surrounds both nuclei. This type of bond is a **covalent bond** and typically develops between elements that are near one another on the periodic table. Bonds between two atoms of the element may be of this type; the bonds in an oxygen molecule (O_2) are a good example. The bonds between carbon and hydrogen

Strong tendency to lose electrons

No tendency to gain or lose electrons

Tendency to share electrons or gain and lose electrons

Strong tendency to gain electrons

Legend:
- Atomic number (protons)
- Ionic charge
- Symbol
- Ionic radius Å
- Name of element

Example: 8 O −2, 1.42, Oxygen

Darker colors are major constituents of crust
Tendency to lose electrons

Metals | Nonmetals

#	Sym	Charge	Radius	Name
1	H			Hydrogen
2	He			Helium
3	Li	1	0.76	Lithium
4	Be	2	0.27	Beryllium
5	B	3	0.11	Boron
6	C	4	0.16	Carbon
7	N	5	0.13	Nitrogen
8	O	−2	1.42	Oxygen
9	F	−1	1.36	Fluorine
10	Ne			Neon
11	Na	1	1.18	Sodium
12	Mg	2	0.72	Magnesium
13	Al	3	0.54	Aluminum
14	Si	4	0.40	Silicon
15	P	5	0.38	Phosphorus
16	S	−2	1.84	Sulfur
17	Cl	−1	1.81	Chlorine
18	Ar			Argon
19	K	1	1.51	Potassium
20	Ca	2	1.12	Calcium
21	Sc	3	0.14	Scandium
22	Ti	4	0.74	Titanium
23	V	3	0.64	Vanadium
24	Cr	3	0.62	Chromium
25	Mn	2	0.83	Manganese
26	Fe	2	0.78	Iron
27	Co	2	0.74	Cobalt
28	Ni	2	0.69	Nickel
29	Cu	2	0.73	Copper
30	Zn	2	0.74	Zinc
31	Ga	3	0.62	Gallium
32	Ge	4	0.53	Germanium
33	As	3	0.46	Arsenic
34	Se	6	0.42	Selenium
35	Br	−1	1.95	Bromine
36	Kr			Krypton
37	Rb	1	1.61	Rubidium
38	Sr	2	1.26	Strontium
39	Y	3	1.02	Yttrium
40	Zr	4	0.84	Zirconium
41	Nb	5	0.74	Niobium
42	Mo	5	0.61	Molybdenum
43	Tc			Technetium
44	Ru	4	0.62	Ruthenium
45	Rh	3	0.67	Rhodium
46	Pd	2	0.86	Palladium
47	Ag	1	1.15	Silver
48	Cd	2	0.95	Cadmium
49	In	3	0.80	Indium
50	Sn	4	0.69	Tin
51	Sb	3	0.76	Antimony
52	Te	6	0.56	Tellurium
53	I	−1	2.16	Iodine
54	Xe			Xenon
55	Cs	1	1.74	Cesium
56	Ba	2	1.42	Barium
57–71				
72	Hf	4	0.83	Hafnium
73	Ta	5	0.74	Tantalum
74	W	6	0.60	Tungsten
75	Re	4	0.63	Rhenium
76	Os	6	0.54	Osmium
77	Ir	4	0.62	Iridium
78	Pt	2	0.60	Platinum
79	Au	1	0.68	Gold
80	Hg	2	1.02	Mercury
81	Tl	1	1.59	Thallium
82	Pb	2	1.29	Lead
83	Bi	3	1.17	Bismuth
84	Po			Polonium
85	At	−1		Astatine
86	Rn			Radon
87	Fr	1		Francium
88	Ra	2	1.48	Radium
89–92				
57	La	3	1.16	Lanthanum
58	Ce	3	1.14	Cerium
59	Pr	3		Praseodymium
60	Nd	3	1.11	Neodymium
61	Pm	3	1.09	Promethium
62	Sm	3	1.08	Samarium
63	Eu	2	1.25	Europium
64	Gd	3	1.05	Gadolinium
65	Tb	3	1.04	Terbium
66	Dy	3	1.03	Dysprosium
67	Ho	3	1.02	Holmium
68	Er	3	1.00	Erbium
69	Tm	3	0.99	Thulium
70	Yb	3	0.98	Ytterbium
71	Lu	3	0.98	Lutetium
89	Ac			Actinium
90	Th	4	1.05	Thorium
91	Pa			Protactinium
92	U	4	1.00	Uranium

FIGURE 3.3 The periodic table of the elements shows the name and symbol of all of the naturally occurring elements. The lightest and simplest elements are in the upper left; across and toward the bottom, each element is progressively more complex, with increasing numbers of nuclear particles and electrons. The elements are separated into rows according to the outermost electron shell. Also shown is the charge of the common ion and the radius for that ion. These properties of an element control how it combines with other elements to form minerals.

in organic materials are also of this type. Many bonds found in natural substances are intermediate between covalent and ionic bonds. Electrons are "pulled" closer to the nucleus of one ion than to the other. As a consequence, one part of the molecule may have a slight charge. The Si-O bond that is so common in minerals is like this.

A third type of bond is the **metallic bond.** In a metal, each atom contributes one or more outer electrons that moves relatively freely throughout the entire aggregate of ions. A given electron is not attached to a specific ion pair but moves about. This sea of negatively charged electrons holds the positive metallic ions together in a crystalline structure and is responsible for the special characteristics of metals, including their high electrical conductivity and ductile behavior. Except for a few native elements (such as gold), few minerals have metallic bonds.

States of Matter

The principal differences between solids, liquids, and gases involve the degree of ordering of the constituent atoms. In the typical **solid,** atoms are arranged in a rigid framework. The arrangement in crystalline solids is quite different. The atomic structure of a crystal consists of a regular, repeating, three-dimensional pattern known as a **crystal structure.** However, there are some **amorphous solids** in which the atomic arrangement is random. Glass is an example of an amorphous solid that lacks a clearly defined crystalline structure. In such solids, each atom occupies a more or less fixed position but has a vibrating motion. Changes in crystalline solids occur as the temperature or pressure changes. For example, as temperature rises, the vibration of atoms in the structure increases, and atoms move farther and farther apart. Eventually the bonds between two atoms may break and they become free and able to glide past one another. Melting ensues, and the crystalline solid passes into the liquid state.

In a **liquid,** the basic particles are in random motion, but they are packed closely together. They slip and glide past one another or collide and rebound, but they

Why can gaseous, liquid, and solid forms of a substance have such different physical properties and still have the same composition?

(A) The formation of an ionic bond in sodium and chloride ions by transfer of an electron from the outermost shell of a sodium atom to the outermost shell of a chlorine atom results in a stable outer shell for each ion.

(B) Covalent bonds form when two atoms share electrons. The bond between silicon and oxygen, so common in minerals, is largely of this type.

FIGURE 3.4   Elements form chemical bonds in several different ways, but all involve interactions of electrons in the outermost electron shell. Ionic and covalent bonds are two of the most important in minerals.

are held together by forces of attraction greater than those in gases. This force of attraction explains why density generally increases and compressibility decreases as matter changes from gas to liquid to solid. If a liquid is heated, the motion of the particles increases, and individual atoms or molecules become separated as they move about at high speeds.

In a **gas,** the particles are in rapid motion and travel in straight lines until their direction is changed by collision. Because the individual atoms or molecules are separated by empty spaces and are comparatively far apart, gases can be markedly compressed and can exert pressure. Gases have the ability to expand indefinitely, and the continuous rapid motion of the particles results in rapid diffusion.

Water undoubtedly provides the most familiar example of matter changing through the three basic states. At pressures prevailing on Earth's surface, water changes from a solid, to a liquid, to a gas in a temperature range of only 100°C. Most people are familiar with the effects of temperature changes on the state of matter because of their experience with water as it freezes, melts, and boils. Fewer people are familiar with the effects of pressure. Under great pressure, water will remain liquid at temperatures as high as 371°C.

The combined effects of temperature and pressure on water are shown in the phase diagram in Figure 3.5. An interesting and very important feature of water is the fact that as it freezes, the solid is actually less dense than the liquid. As a result, water ice floats rather than sinks. The expansion of water during freezing is important for weathering and in the moderation of Earth's climate. Because polar ice floats on the sea, it creates an insulating layer that slows the cooling of the rest of the sea. If ice did not float, Earth's oceans may have frozen solid during the ice ages.

Other forms of matter in the solid Earth are capable of similar changes, but usually their transitions from solid, to liquid, to gas occur at comparatively high temperatures. At normal room temperature and pressure, 93 of the 106 elements are solids, 2 are liquids, and 11 are gases. Diagrams similar to Figure 3.5, constructed from laboratory work on other minerals, provide important insight into the processes operating at the high temperatures and pressures below Earth's surface.
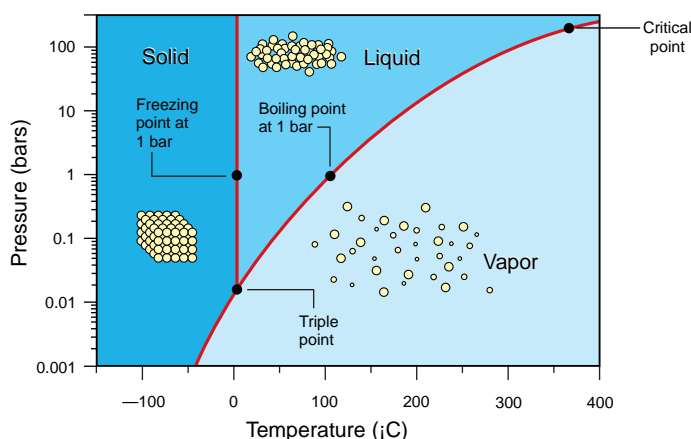
FIGURE 3.5 **Temperature and pressure** determine the state in which matter exists. In this diagram, the ranges of temperature and pressure for the various phases of water are shown. The triple point is the point at which all three phases are in equilibrium. Beyond the critical point the liquid and gas phases cannot be distinguished. Similar phase diagrams can be constructed for other minerals.

THE NATURE OF MINERALS

> A mineral is a natural inorganic solid with a specific internal structure and a chemical composition that varies only within specific limits. All specimens of a given mineral, regardless of where, when, or how they were formed, have the same physical properties (including cleavage, crystal form, hardness, density, color, luster, and streak). Minerals also have restricted stability ranges.

Minerals are the solid constituents of Earth. Many people think of minerals only as exotic crystals in museums or as valuable gems and metals; but grains of sand, snowflakes, and salt particles are also minerals, and they have much in common with gold and diamonds. A precise definition is difficult to formulate, but for a substance to be considered a **mineral,** it must meet the conditions listed above and described in greater detail below. The differences among minerals arise from the kinds of atoms they contain and the ways those atoms are arranged in a crystalline structure.

Natural Inorganic Solids

By definition, only naturally occurring inorganic solids are minerals—that is, natural elements or inorganic compounds in a solid state. Synthetic products, such as artificial diamonds, are therefore not minerals in the strict sense. Organic compounds, such as coal and petroleum, which lack a crystal structure, are also not considered to be minerals. This criterion is not as important as most of the others. After all, there is little difference between a synthetic and a natural gem, other than where they formed. All of its structural, physical, and chemical properties are shared with its natural counterparts. Likewise, there are organic solids that have all of the characteristics of minerals.

Why is the structure of a mineral so important?

The Structure of Minerals

The key words in the definition of mineral are *internal structure.* Minerals can consist of a single element, such as gold, silver, copper, diamond, or sulfur. However, most are compounds of two or more elements. The component atoms of a mineral have a specific arrangement in a definite geometric pattern. All specimens of a given mineral have the same internal structure, regardless of when, where, and how they were formed. This property of minerals was suspected

With modern methods of **X-ray diffraction,** we can determine precisely a mineral's internal structure and learn much about the arrangement of its atoms. Diffraction involves the bending of X rays as they pass through a crystalline substance.

The technique is illustrated in the figure below. When a narrow beam of X rays is passed through a mineral grain, the X rays are diffracted by the framework of atoms. The individual ions are spaced very closely in the rigid network, close enough to bend X rays—like a diffraction grating bends light rays. The diffracted rays cause constructive and destructive interference—in effect, concentrating the energy of the X rays in some areas and dispersing it in others. After they leave the crystal, the X rays expose a photographic plate or are detected with a scanning device and plotted as shown. From the pattern made by the spots or from measurements of the position and height of the peaks, the systematic orientation of planes of atoms within the crystal can be deduced. Such measurements are so precise that the distances between atoms can be measured and the size and shape of the electron cloud calculated. Detailed models of crystal structures showing the position of each different atom can thus be constructed.

The X-ray diffraction instrument is now the most basic device for determining the internal structure of minerals, and geologists use it extensively for precise mineral identification and analysis.

Two typical examples of X-ray spectra are shown on the chart. The lower curve is the X-ray diffraction pattern for the mineral quartz. The peaks are created by constructive interference of the X rays and correspond to specific atomic spacings that are the result of the nearly covalent silicon-oxygen bond. The peak positions do not directly reveal the kinds of atoms, only their distances and arrangements. The more complicated X-ray diffraction pattern was formed from a specimen of feldspar. Because feldspars have a much greater variety of elements, bond types, and structural elements, the diffraction pattern is also more complicated.

X-ray diffraction analysis is the definitive technique that shows us that each mineral species has its own distinctive structure that is repeated many times in every grain of the mineral. It reveals the great symmetry and order found in the mineral kingdom.
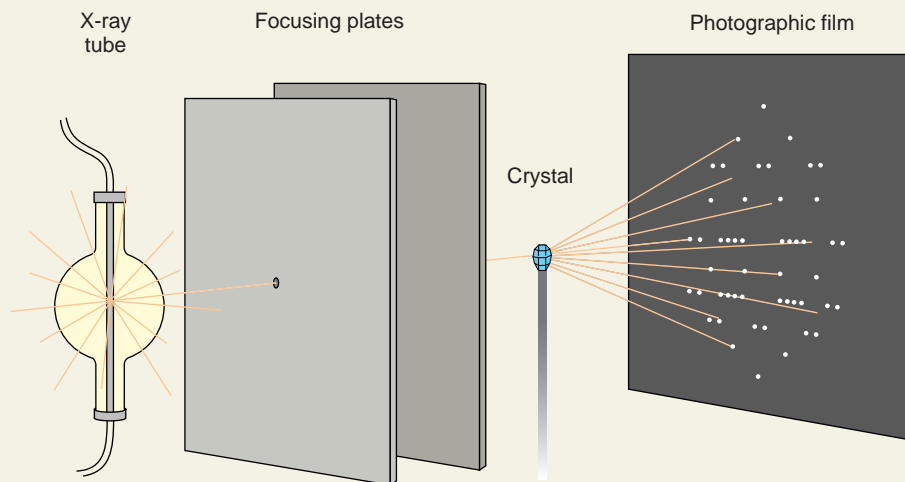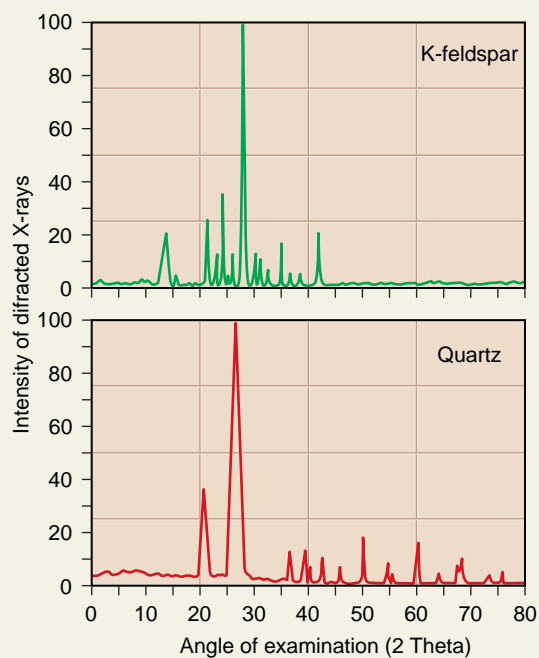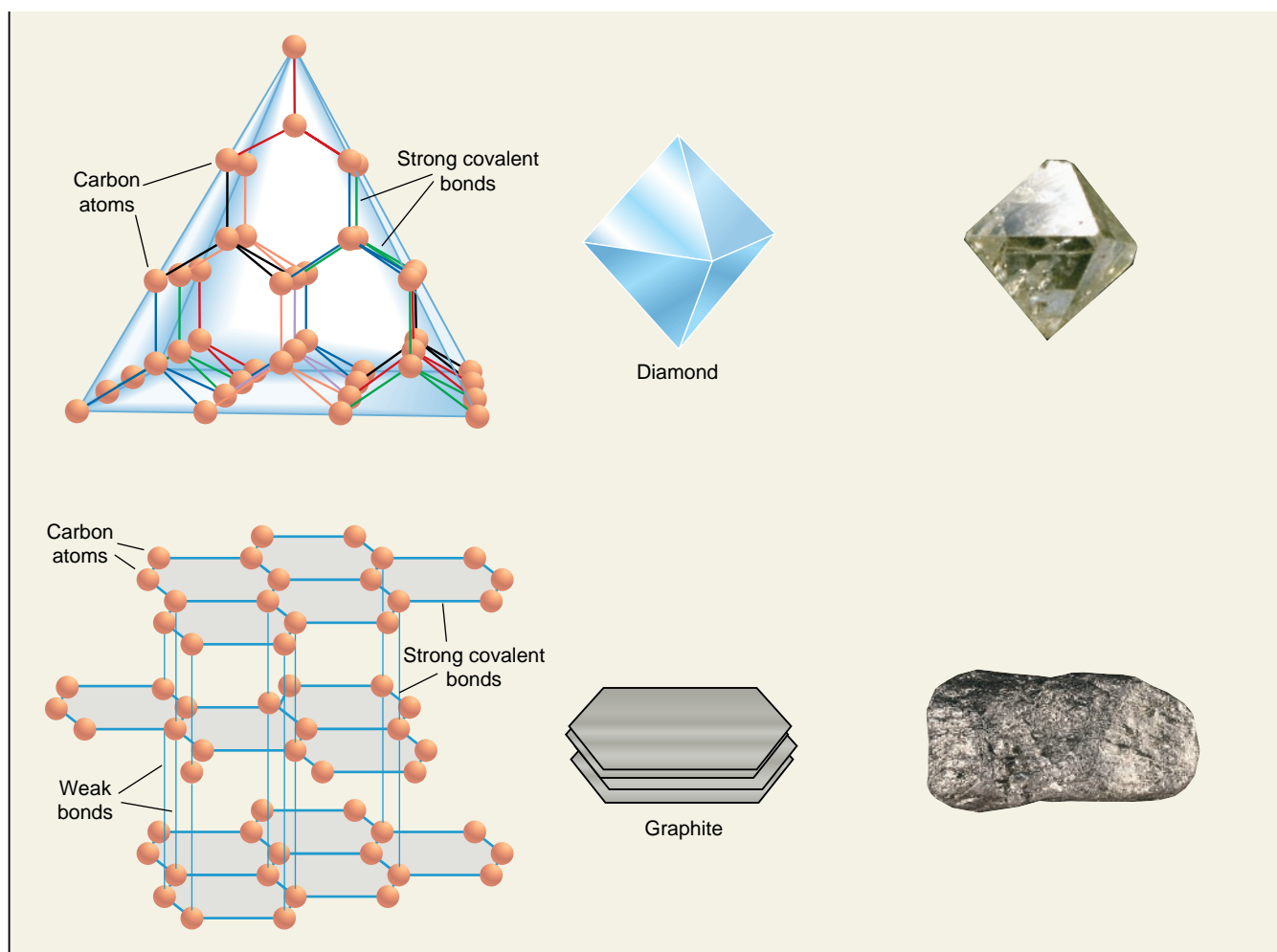
FIGURE 3.6   The internal structure of a mineral controls its physical properties. Diamond and graphite have exactly the same chemical composition, but the carbon atoms are arranged differently and held together by different types of bonds. Graphite is made of sheets of carbon stacked on top of one another. It is soft and black. Diamond, the hardest mineral known, is made of carbon atoms bound together in a tight tetrahedral framework. Most grains of diamond are transparent. *(Photographs by Jeffrey A. Scovil)*

long ago by mineralogists who observed the many expressions of order in **crystals.** Nicolaus Steno (1638–1687), a Danish monk, was among the first to note this property. He found from numerous measurements that each of the different kinds of minerals has a characteristic crystal form. Although the size or shape of a mineral's crystalline form may vary, similar pairs of crystal faces always meet at the same angle. This is known as the *law of constancy of interfacial angles.*

Later, Rene Hauy (1743–1822), a French mineralogist, accidentally dropped a large crystal of calcite and observed that it broke along three sets of planes only, so all the fragments had a similar shape (see Figure 3.9). He then proceeded to break other calcite crystals in his own collection, plus many in the collections of his friends, and found that all of the specimens broke in exactly the same manner. All of the fragments, however small, had the shape of a rhombohedron. To explain his observations, he assumed that calcite is built of innumerable infinitely small rhombohedra packed together in an orderly manner; he concluded that the cleavage of calcite is related to the ease of parting of such units from adjacent layers. His discovery was a remarkable advance in understanding crystals. Today we know that cleavage planes are planes of weakness in the crystal structure and that they are not necessarily parallel to the crystal faces. Cleavage planes do, however, constitute a striking expression of the orderly internal structure of crystals.

To understand the importance of structure in a mineral, consider the characteristics of diamond and graphite (Figure 3.6). These two minerals are identical in

Si^{+4} 0.40   Al^{+3} 0.54

Ni^{+2} 0.69   Mg^{+2} 0.72   Fe^{+2} 0.78   Fe^{+3} 0.64

K$^+$ 1.51   Na$^+$ 1.18   Ca^{+2} 1.12

O^{-2} 1.42   OH^{-1} 1.40
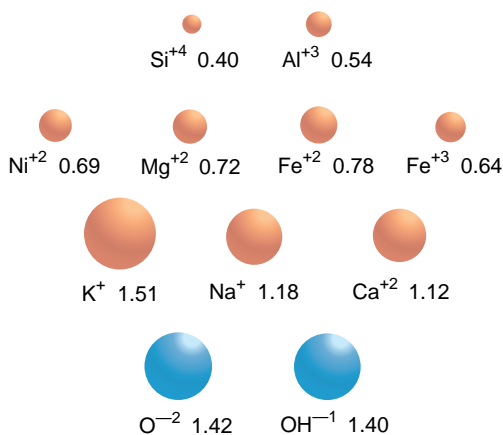
FIGURE 3.7   The relative size and electrical charge of ions are important factors governing the suitability of one ion to substitute for another in a crystal structure. Silicon can be replaced by aluminum, iron by magnesium or nickel, and sodium by calcium.

chemical composition. Both consist of a single element, carbon (C). Their crystal structures and physical properties, however, are very different. In diamond, which forms only under high pressure, the carbon atoms are packed closely, and the covalent bonds between the atoms are very strong. Their structure explains why diamonds are extremely hard—the hardest natural substance known. In graphite, the carbon atoms form layers that are loosely bound. Because of weak bonds, the layers separate easily, so graphite is slippery and flaky. Because of its softness and slipperiness, graphite is used as a lubricant and is also the main constituent of common "lead" pencils. The important point to note is that different structural arrangements of exactly the same elements produce different minerals with different properties. This ability of a specific chemical substance to crystallize in more than one type of structure is known as **polymorphism.**

The Composition of Minerals

A mineral has a definite chemical composition, in which specific elements occur in definite proportions. Thus, a precise chemical formula can be written to express the chemical composition—for example, SiO_2, $CaCO_3$, and so on. The chemical composition of some minerals can vary, but only within specific limits. In these minerals, two or more kinds of ions can substitute for each other in the mineral structure, a process known as **ionic substitution.** Ionic substitution results in a chemical change in the mineral without a change in the crystal structure, so substitution can occur only within definite limits. The composition of such a mineral can be expressed by a chemical formula that specifies ionic substitution and how the composition can change.

The suitability of one ion to substitute for another is determined by several factors, the most important being the size and the electrical charge of the ions in question (Figures 3.3 and 3.7). Ions can readily substitute for one another if their ionic radii differ by less than 15%. If a substituting ion differs in charge from the ion for which it is substituted, the charge difference must be compensated for by other substitutions in the same structure in order to maintain electrical neutrality.

Ionic substitution is somewhat analogous to substituting different types of equal-sized bricks in a wall. The substitute brick may be composed of glass, plastic, or whatever, but because it is the same size as the original brick, the structure of the wall is not affected. An important change in composition has, however, occurred, and as a result there are changes in physical properties. In minerals, ionic substitution causes changes in hardness and color, for example, without changing the internal structure.

Ionic substitution is common in rock-forming minerals and is responsible for mineral groups, the members of which have the same structure but varying composition. For example, in the olivine group, with the formula $(Mg, Fe)_2SiO_4$, ions of iron (Fe^{+2}) and magnesium (Mg^{+2}) can substitute freely for one another because they have similar charges and sizes (Figure 3.7). The total number of Fe^{+2} and Mg^{+2} ions is constant in relation to the number of silicon (Si) and oxygen (O) atoms in the olivine, but the ratio of iron to magnesium may vary in different samples. The common minerals feldspar, pyroxene, amphibole, and mica each constitute a group of related minerals in which ionic substitution produces a range of chemical composition.

The Physical Properties of Minerals

What determines the physical properties of a mineral?

Because a mineral has a definite chemical composition and internal crystalline structure, all specimens of a given mineral, regardless of when or where they were formed, have the same physical and chemical properties. If ionic substitution occurs, variation in physical properties also occurs, but because ionic substitution can occur only within specific limits, the range in physical properties also can occur only within specific limits. This means that one piece of quartz, for example,

(A) Prismatic tourmaline
$[Na(Li,Al)_3Al_6(BO_3)_3Si_6O_{18}(OH)_4]$.
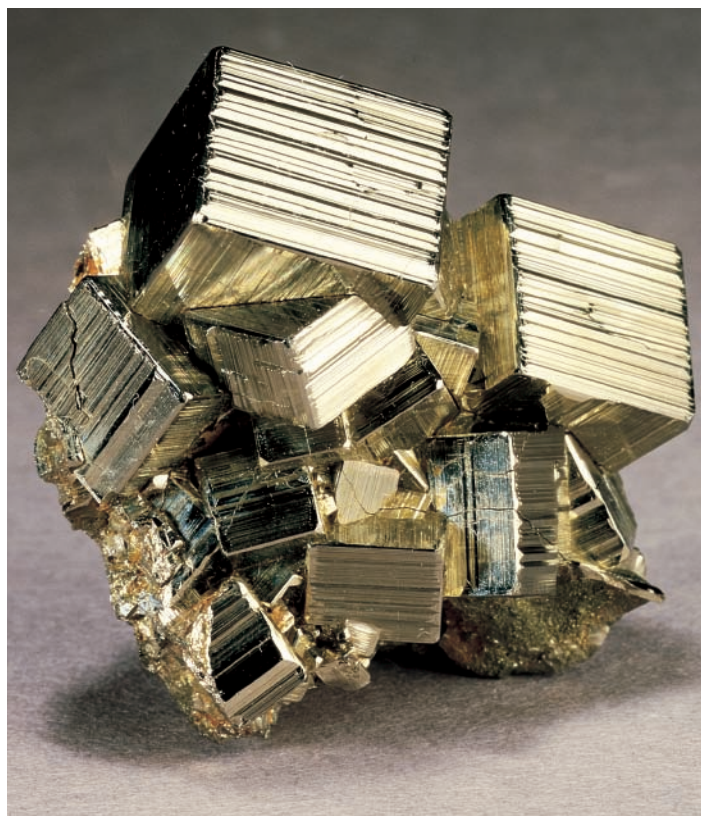

(B) Tetrahedrons of sphalerite (ZnS).

(C) Needles of the rare mineral crocoite ($PbCrO_4$).

(D) Radiating clusters of long slender needles of the zeolite
mineral mordenite $(Ca,Na_2,K_2)(Al_2Si_{10})(O_{24}\cdot7H_2O)$.

FIGURE 3.8   **Crystal form** is an important physical
property showing the arrangement of atoms in a mineral.
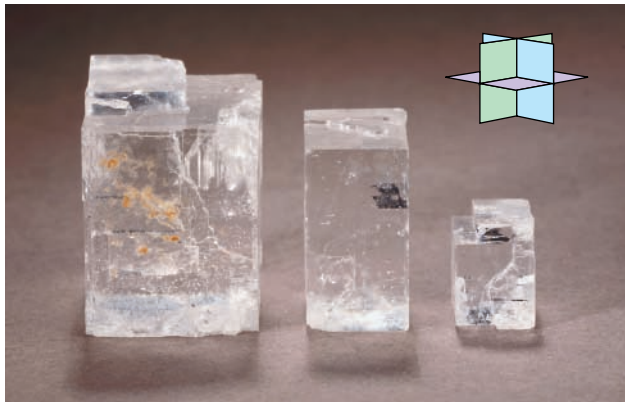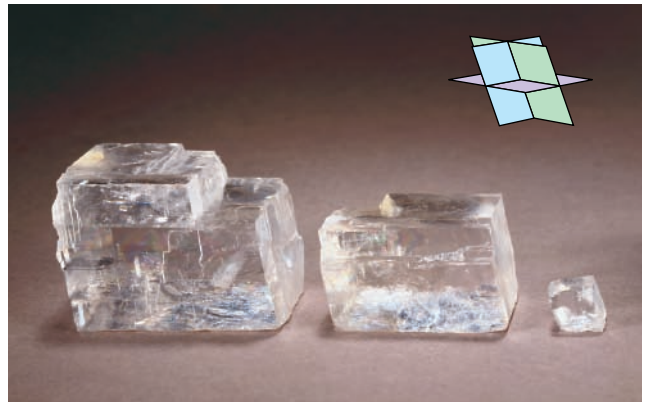*(Photographs by Jeffrey A. Scovil)*

(E) Cubes of pyrite (FeS_2), commonly known as fool's gold.

(A) One plane of cleavage in mica produces thin plates or sheets.
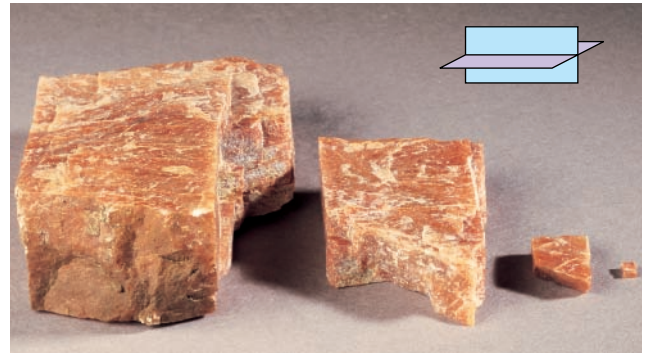
(B) Two planes of cleavage at right angles in feldspar produce blocky fragments.

(C) Three planes of cleavage at right angles in halite produce cubic fragments.

(D) Cleavage of calcite occurs in three planes that do not intersect at right angles, forming rhombohedrons.

FIGURE 3.9   **Cleavage** reflects planes of weakness within a crystal structure.

is as hard as any other piece, that it has the same density, and that it breaks in the same manner, regardless of when, where, or how it was formed.

   The more significant and readily observable physical properties of minerals are crystal form, cleavage, hardness, density, color, luster, and streak.

   If a crystal is allowed to grow in an unrestricted environment, it develops natural **crystal faces** and assumes a specific geometric **crystal form.** The shape of a crystal is a reflection of the internal structure and is an identifying characteristic for many mineral specimens (Figure 3.8). If the atoms are arranged in a long chain, the crystal may be shaped like a needle. If the atoms are arranged in a boxlike network, the crystal will likely be in the form of a cube. If the space for growth is restricted, however, smooth crystal faces cannot develop.

   Cleavage is the tendency of a crystalline substance to split or break along smooth planes parallel to zones of weak bonding in the crystal structure (Figure 3.9). If the bonds are especially weak in a given plane, as in graphite, mica, or halite, perfect cleavage occurs with ease. Breaking the mineral in any direction other than along a cleavage plane is difficult (Figure 3.9). In other minerals, the differences in bond strength are not great, so cleavage is poor or imperfect. Cleavage can occur in more than one direction, but the number and direction of cleavage planes in a given mineral species are always the same. Some minerals have no weak planes in their crystalline structure, so they do not have cleavage and break along various types of fracture surfaces. Quartz, for example, characteristically breaks by **conchoidal fracture**—that is, along curved surfaces, like the curved surfaces of chipped glass. Cleavage planes and crystal faces should not be confused with the facets found on gems. Facets are produced by grinding and polishing the surface of a

FIGURE 3.10   Hardness reflects the strength of the atomic bonds inside the mineral. Gypsum has a hardness of 2 on Mohs hardness scale. It is a very soft mineral and can easily be scratched with a fingernail.

TABLE 3.1

Mohs Hardness Scale

Hardness	Mineral	Test
1	Talc	
2	Gypsum	
		Fingernail
3	Calcite	
		Copper coin
4	Fluorite	
5	Apatite	
		Knife blade or glass plate
6	K-feldspar	
7	Quartz	
		Steel file
8	Topaz	
9	Corundum	
10	Diamond	

mineral grain and do not necessarily correspond to cleavage directions. For example, diamond lacks cleavage altogether but can be polished so that a single crystal will have many shiny faces.

Hardness is a measure of a mineral's resistance to abrasion. It is in effect a measure of the strength of the atomic bonds in a crystal. This property is easily determined and is used widely for field identification of minerals. More than a century ago, Friedrich Mohs (1773–1839), a German mineralogist, assigned arbitrary relative numbers to 10 common minerals in order of their hardness. He assigned the number 10 to diamond, the hardest mineral known. Softer minerals were ranked in descending order, with talc, the softest mineral, assigned the number 1. The Mohs hardness scale (Table 3.1) provides a standard for testing minerals for preliminary identification. Gypsum, for example, has a hardness of 2 and can be scratched by a fingernail (Figure 3.10). More exacting measures of hardness show that diamond is by far the hardest mineral.

Density is the ratio of the weight of a substance to its volume. For example, at room temperature, 1 cm^3 of water weighs 1 g; the density is thus 1 g/cm^3. On the other hand, 1 cm^3 of solid lead weighs a little over 11 g, and thus its density is 11 g/cm^3.

Density is one of the more precisely defined properties of a mineral. It depends on the kinds of atoms making up the mineral and how closely they are packed in the crystal structure. Clearly, the more numerous and compact the atoms, the higher the density. Most common rock-forming minerals have densities that range from 2.65 g/cm^3 (for quartz) to about 3.37 g/cm^3 (for magnesium olivine). Iron-rich olivine is even denser (4.4 g/cm^3) because iron has a higher atomic weight than magnesium. Some metallic minerals have much higher densities. For example, native gold has a density of about 20 g/cm^3 and native iron has a density of almost 8 g/cm^3. At high pressures, the densities of most minerals increase because the atoms are forced to be closer together. At high temperatures, their densities decrease as the atoms move farther apart.

Color is one of the more obvious properties of a mineral. Unfortunately, it is not diagnostic. Most minerals are found in various hues, depending on such factors as subtle variations in composition and the presence of inclusions and impurities. Quartz, for example, ranges through the spectrum from clear, colorless crystals to purple, red, white, yellow, gray, and black.

Luster describes the appearance of light reflected from a mineral's surface. Luster is described only in subjective, imprecise terms. There are two basic kinds of luster: metallic and nonmetallic. Minerals with a metallic luster shine like metals. Nonmetallic luster ranges widely, including vitreous (glassy), porcelainous, resinous,

FIGURE 3.11   The stable form of SiO₂ depends on pressure and temperature. The colored areas show the range of temperature and pressure over which each of five different minerals are stable. Quartz, for example, is stable at intermediate temperatures over a wide range of pressures. Other minerals of the same composition (SiO₂), but with different atomic arrangements, are stable at other pressures and temperatures.

and earthy (dull). The luster of a mineral is controlled by the kinds of atoms and by the kinds of bonds that link the atoms together. Many minerals with covalent bonds have a brilliantly shiny luster, called adamantine luster, as in diamond. Ionic bonds create more vitreous luster, as in quartz. Metallic bonding in native metals, such as gold, also has its characteristic luster.

Streak refers to the color of a mineral in powder form and is usually more diagnostic than the color of a large specimen. For example, the mineral pyrite (fool's gold) has a gold color but a black streak, whereas real gold has a gold streak—the same color as that of larger grains. Streak is tested by rubbing a mineral vigorously against the surface of an unglazed piece of white porcelain. Minerals softer than the porcelain leave a streak, or line, of fine powder. For minerals harder than porcelain, a fine powder can be made by crushing a mineral fragment. The powder is then examined against a white background.

Magnetism is a natural characteristic of only a few minerals, like the common iron oxide magnetite. Although only a few minerals can be identified using this property, magnetism is an important physical property of rocks that is used in many investigations of how Earth works (see page 604).

Stability Ranges

Another important feature of each mineral is that it is stable only over a fixed range of conditions. We call a mineral **stable** if it exists in equilibrium with its environment. In such a case, there is little tendency for further change. The environment that exists when a mineral crystallizes determines which of the many thousands of minerals will form. The environmental conditions that determine whether a particular mineral is stable are mainly pressure, temperature, and composition.

We have already examined the **stability ranges** for the various states of water (see Figure 3.5), and we can use similar phase diagrams to represent the range of conditions over which a specific mineral is stable. Figure 3.11 shows the names and stability fields for various minerals with the chemical formula SiO_2. Quartz is the most common of these minerals because it is stable over the range of temperatures and pressures found near Earth's surface. However, if the temperature is increased to 1300°C at a pressure of 1000 bars (a depth in Earth of about 3 km), the arrangement of the atoms in quartz will change to form a different mineral called tridymite, which has its own structure and distinctive physical properties. For example, quartz has a density of 2.65 g/cm^3 and tridymite has a density of about 2.26 g/cm^3. If the temperature is increased to 1600°C, still another change occurs as tridymite converts to cristobalite with a density of 2.33 g/cm^3. In the absence of water, pure SiO_2 melts only at a temperature higher than 1700°C. Changes in pressure can also induce minerals to break down and form new species that are stable under the new conditions. Metamorphic processes, discussed in more detail in Chapter 6, are driven by the tendency for minerals to react and change as their environment changes.

Although minerals have distinctive stability ranges, they may remain in existence far from those conditions. A mineral existing outside of its stability range is called **metastable.** Metastability occurs if the reactions to form new minerals from preexisting minerals are very slow. Such barriers are common at Earth's surface, where low temperatures make atomic movements and reactions very sluggish in solids. Thus, tridymite has been found at temperatures far below the range of temperatures shown in Figure 3.11. Moreover, feldspars are common at Earth's surface, even though clay minerals are more stable in the presence of water. Despite these reaction barriers at low temperature, it is useful to keep in mind the approximate range of temperatures and pressures over which a given mineral is stable.

THE GROWTH AND DESTRUCTION OF MINERALS

> Minerals grow as matter changes from a gaseous or liquid state to a solid state or when one solid recrystallizes to form another. They break down as the solid changes back to a liquid or a gas. All minerals came into being because of specific physical and chemical conditions, and all are subject to change as these conditions change. Minerals, therefore, are an important means of interpreting the changes that have occurred in Earth throughout its history.
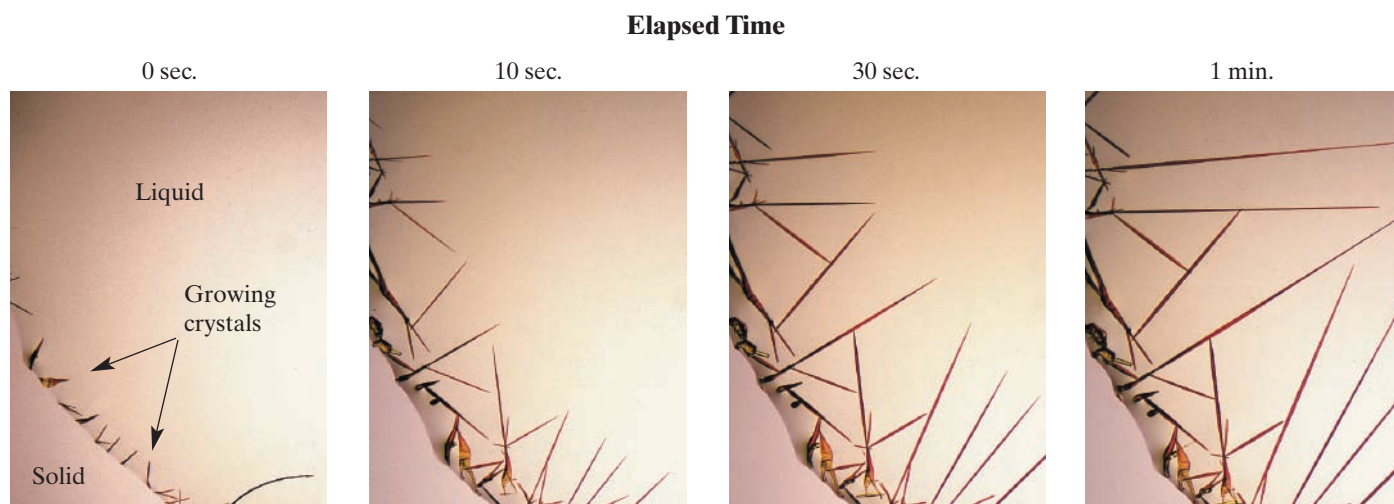
Crystal Growth

Even though minerals are inorganic, they can grow. Growth is accomplished by **crystallization,** which occurs by the addition of ions to a crystal face. As noted above, an environment suitable for crystal growth includes (1) proper concentration of the kinds of atoms or ions required for a particular mineral and (2) proper temperature and pressure.

The time-lapse photographs in Figure 3.12 show how crystals grows from a liquid in an unrestricted environment. Although the size of each crystal increases, its form and internal structure remain the same. New atoms are added to the faces of the crystal, parallel to the plane of atoms in the basic structure. Some crystal faces, however, grow faster than others. As a result of these different growth rates, the crystal may become elongated in one direction. Thus, the ideal crystal shape reflects not only the arrangement of atoms inside the crystal, but it also controlled by which faces grow faster or slower. You can see that all of the crystals in Figure 3.12 have the same idealized shape, because they are all the same mineral. The mineral grains in the chapter opening photograph show the dramatic results of growth in an unrestricted environment. These minerals crystallized from a watery solution in an open vug within an ancient series of lava flows. Each crystal was free to grow to its ideal shape with little interference from other crystals. It is easy to tell that there are two different kinds of minerals from their ideal shapes.

In contrast, where space is restricted, a crystal may not grow to form its ideal crystal shape. Where a growing crystal encounters a barrier (such as another crystal), it simply stops growing. This process is illustrated in Figure 3.12. Note how the vertical crystal grew between 10 and 30 seconds. At 30 seconds, it has impinged on a nearly horizontal crystal and stopped growing. However, the more horizontal

How can a mineral, which is inorganic, grow?

Elapsed Time

0 sec.	10 sec.	30 sec.	1 min.

FIGURE 3.12   Crystal growth can be recorded by time-lapse photography. Each crystal grows as atoms in the surrounding liquid lock onto the outer faces of the crystal structure.
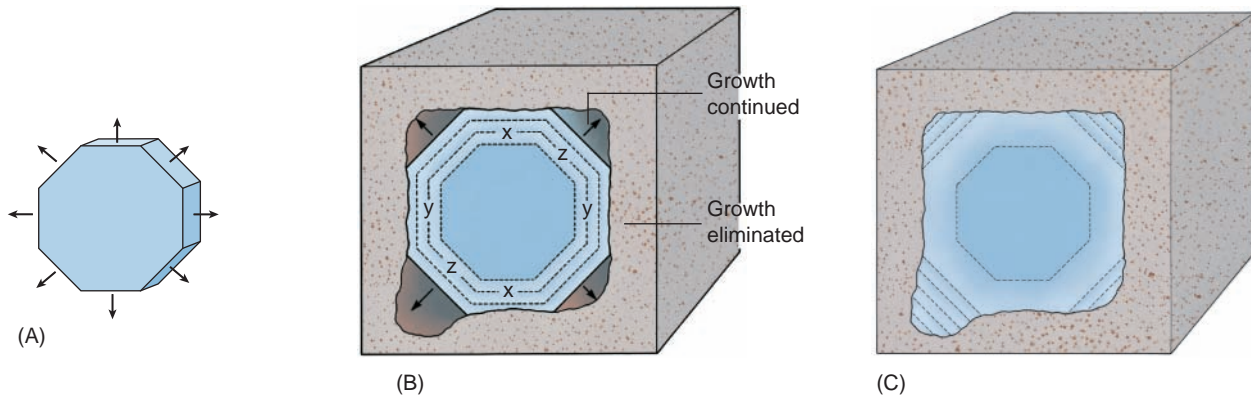
Figure 3.13 **Crystals growing in a restricted environment** do not develop perfect crystal faces. **(A)** Where growth is unrestricted, all crystal faces grow with equal facility. **(B)** In a restricted environment, growth on certain crystal faces, such as *x* and *y*, is terminated but growth on the faces labeled *z* continues. **(C)** The final shape of the crystal is determined by the geometry of the available space in which it grows.

crystal grows throughout the sequence because there were no restrictions to its growth.

Figure 3.13 shows how crystal growth occurs in a restricted environment. A crystal growing from a liquid in a restricted space assumes the shape of the confining area, and well-developed crystal faces do not form. The external form of the crystal can thus take on practically any shape, but its internal structure is in no way modified. The mineral's internal structure remains the same; its composition is unaffected, and no changes in its physical and chemical properties occur. The only modification is a change in the shape of the crystal.

Crystal growth in restricted spaces is common for rock-forming minerals. In a still molten lava flow or in an aqueous solution, many crystals grow at the same time and must compete for space. As a result, in the later stages of growth, crystals in rocks commonly lack well-defined crystal faces and typically interlock with adjacent crystals to form a strong, coherent mass (Figure 3.14). This interlocking texture is especially common in igneous rocks, which form by crystallization from molten rock material.

Most crystals are rather small, measuring from a few tenths of a millimeter to several centimeters in diameter. Some are so small they can be seen only when enlarged thousands of times with a high-powered electron microscope (Figure 3.15). Where crystallization occurs from a mobile fluid in an unrestricted environment, however, crystals can grow to enormous sizes (Figure 3.16).

Destruction of Crystals

Mineral grains can be destroyed in many different ways. Minerals **melt** by removal of outer atoms from the crystal structure as they enter a less organized liquid state. The heat that causes a crystal to melt increases atomic vibrations enough to break
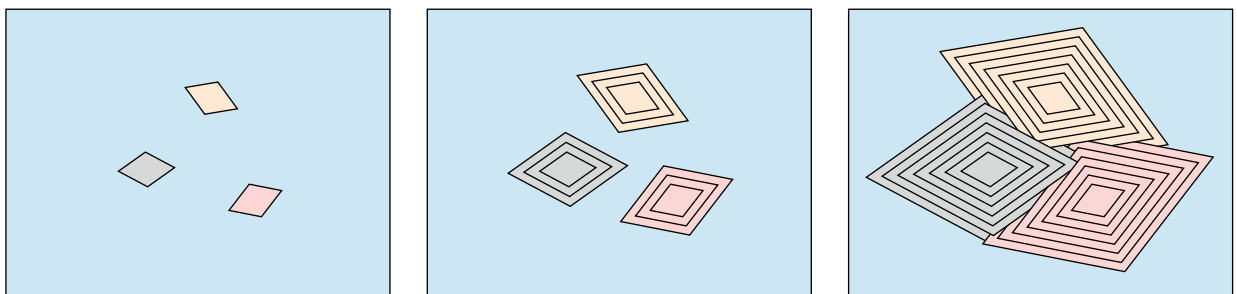
FIGURE 3.14 **Interlocking texture** develops if crystals grow in a restricted environment. Crystals grow into one another when they are forced to compete for space. Such textures are common in igneous rocks which form form molten magma. (see Chapter 4).
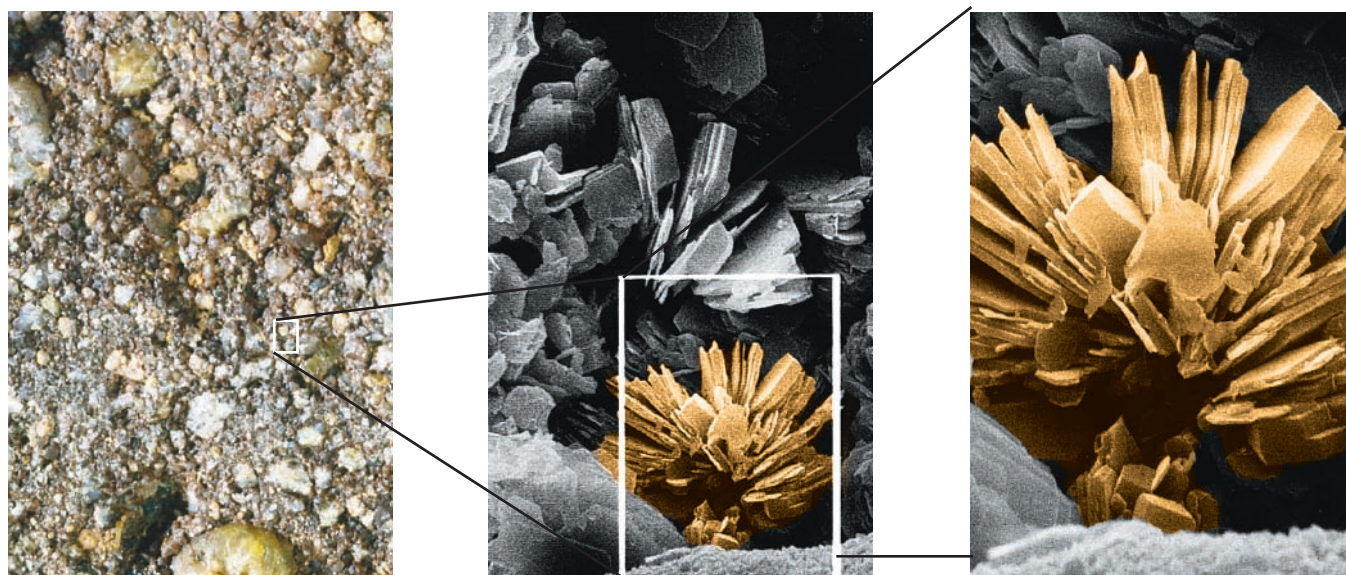
(A) Sand grains magnified 50 times. Small crystals of clay form between the grains.

(B) Clay crystals coating sand grains magnified 1000 times.

(C) Clay crystals magnified 2000 times.

FIGURE 3.15   Submicroscopic crystals of hexagonal plates of clay growing in the pore spaces between sand grains can be seen with an electron microscope. Each crystal contains all of the physical and chemical properties of the mineral, even though each one is extremely small. *(Courtesy of Harry W. Fowkes)*

the bonds holding an atom to the crystal structure. Similarly, atoms can be "pried" loose and carried away by a solvent, usually (in geologic processes) water. Crystals begin to break down or dissolve at the surface and the reaction moves inward.

Mineral grains can also be destroyed as their constituent atoms become rearranged in the solid state. Such **recrystallization** processes are especially common deep inside the crust and mantle, where heat and pressure cause some crystal structures to collapse and new minerals, with a denser, more compact atomic structure (Figure 3.17) to form in their place. In this case, the atoms do not move far, but new bonds form and new internal structures are created. The new mineral grains have different physical properties, like cleavage, luster, hardness, and density.
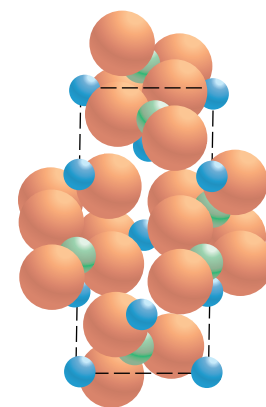
(A) Open structure at low pressure.
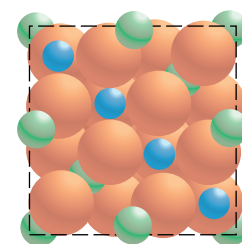
(B) Densely packed structure at high pressure.

FIGURE 3.17   Under high pressure, the atomic structure of a mineral can collapse into a denser form, in which the atoms are more closely packed. Although the physical properties change, the chemical composition may remain the same.

FIGURE 3.16   Large crystals can form where there is ample space for growth, as in caves. These crystals of gypsum are more than 1 m long.

TABLE 3.2

Concentrations of the Most Abundant Elements in Earth's Crust (by weight)

Element	Percentage
O	46.60
Si	27.72
Al	8.13
Fe	5.00
Ca	3.63
Na	2.83
K	2.59
Mg	2.09
Ti	0.44
H	0.14
P	0.12
Mn	0.10
S	0.05
C	0.03

After B. Mason and C. B. Moore, *Principles of Geochemistry,* 4th ed. (New York: Wiley, 1982).

Crystal Models

SILICATE MINERALS

More than 95% of Earth's crust is composed of silicate minerals, a group of minerals containing silicon and oxygen linked in tetrahedral units, with four oxygen atoms to one silicon atom. Several fundamental configurations of tetrahedral groupings are single chains, double chains, two-dimensional sheets, and three-dimensional frameworks.

Although more than 4000 minerals have been identified, 95% of the volume of Earth's crust is composed of a group of minerals called the silicates. This should not be surprising because silicon and oxygen constitute nearly three-fourths of the mass of Earth's crust (Table 3.2) and therefore must predominate in most rock-forming minerals. Silicate minerals are complex in both chemistry and crystal structure, but all contain a basic building block called the silicon-oxygen tetrahedron. Nearly covalent Si-O bonds form a complex ion $[(SiO_4)^{4-}]$ in which four large oxygen ions (O^{2+}) are arranged to form a four-sided pyramid with a smaller silicon ion (Si^{4+}) bonded between them (Figure 3.18). This geometric shape is known as a tetrahedron. The major groups of silicate minerals differ mainly in the arrangement of such silicate tetrahedrons in their crystal structures.

Perhaps the best way to understand the unifying characteristics of the **silicates,** as well as the reasons for the differences, is to study the models shown in Figure 3.19. These were constructed on the basis of X-ray studies of silicate crystals. **Silicon-oxygen tetrahedrons** combine to form minerals in two ways. In the simplest combination, the oxygen ions of the tetrahedrons form bonds with other elements, such as iron or magnesium. Olivine is an example. Most silicate minerals, however, are formed by the sharing of an oxygen ion between two adjacent tetrahedrons. In this way, the tetrahedrons form a larger ionic unit, just as beads are joined to form a necklace. The sharing of oxygen ions by the silicon ions results in several fundamental configurations of tetrahedral groups. These structures define the major silicate mineral groups:

1. Isolated tetrahedrons (example: olivine)
2. Single chains (example: pyroxene)
3. Double chains (example: amphibole)
4. Two-dimensional sheets (examples: micas, chlorite, and clays)
5. Three-dimensional frameworks (examples: feldspars and quartz)

The unmatched electrons of the silicate tetrahedron are balanced by various metal ions, such as ions of calcium, sodium, potassium, magnesium, and iron. The silicate minerals thus contain silicon-oxygen tetrahedrons linked in various patterns by metal ions. Considerable ionic substitution can occur in the crystal structure. For example, sodium can substitute for calcium, or iron can substitute for magnesium. Minerals of a major silicate group can thus differ chemically from one another but have a common silicate structure.
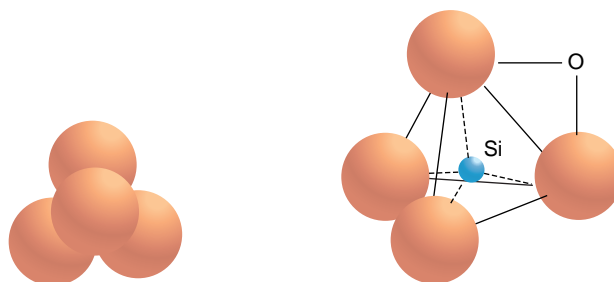
FIGURE 3.18   The silicon-oxygen tetrahedron is the basic building block of the silicate minerals. In this figure, the diagram on the right is expanded to show the position of the small silicon atom. Four large oxygen ions are arranged in the form of a pyramid (tetrahedron), with a small silicon ion covalently bonded into the central space between them. This is the most important building block in geology because it is the basic unit for 95% of the minerals in Earth's crust.

Isolated   Single chain   Double chain

Two-dimensional sheet   Three-dimensional framework
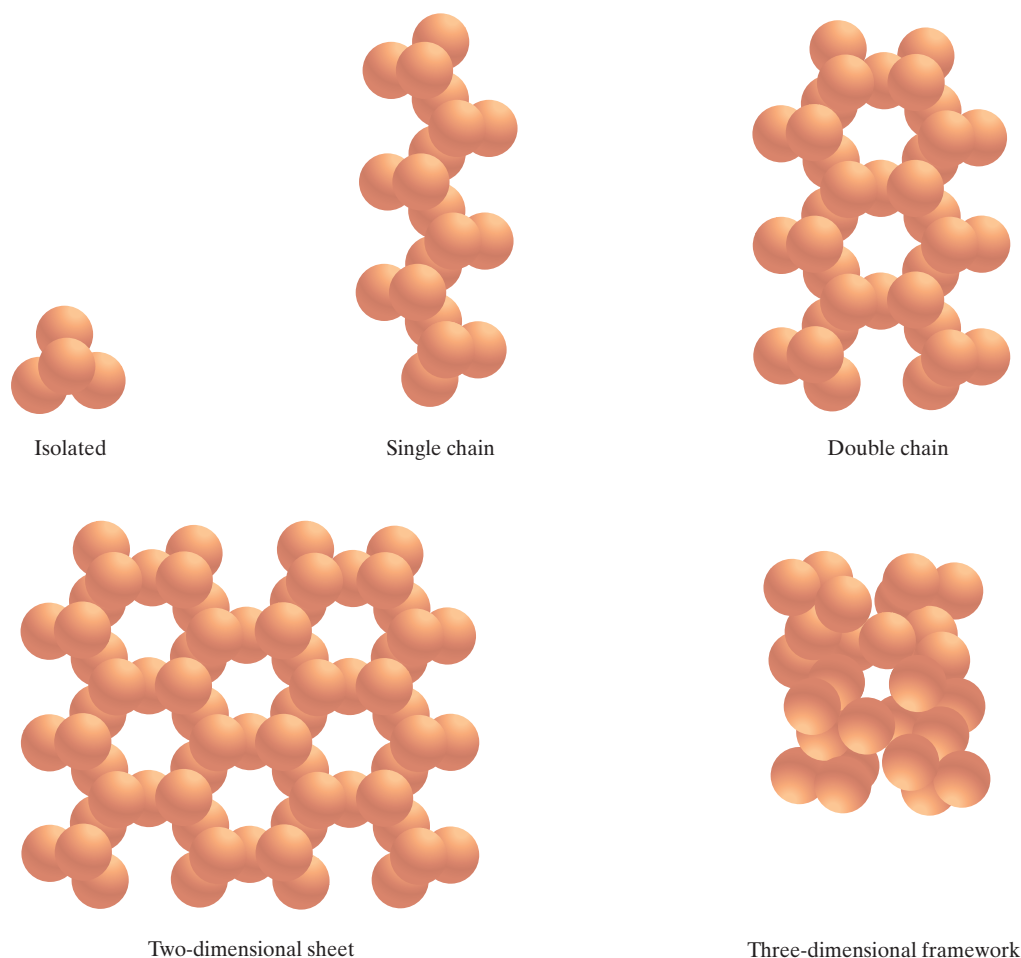
FIGURE 3.19   Silicon-oxygen tetrahedral groups can form various structures by the sharing of oxygen ions among silicon ions. A small silicon ion lies at the center of each tetrahedral unit. In general, various types of metal ions complete the mineral structure; they are not shown here.

ROCK-FORMING MINERALS

Fewer than 20 kinds of minerals account for the great bulk of Earth's crust and upper mantle. The most common silicate minerals are feldspars, quartz, micas, olivine, pyroxenes, amphiboles, and clay minerals. Important nonsilicates are calcite, dolomite, halite, and gypsum.

Most of Earth's crust and upper mantle are composed of silicate minerals in which the common elements—such as iron, magnesium, sodium, calcium, potassium, and aluminum—combine with silicon and oxygen. The identification of these minerals presents some special problems. Rock-forming minerals rarely have well-developed *crystal faces* because (1) they grow by crystallization from melts (e.g., magmas) or from aqueous solutions (e.g., seawater) and vigorously compete for space; (2) they are abraded as they are transported as sediment; or (3) they are deformed under high temperature and pressure. In addition, most rock-forming mineral grains are small, generally less than the size of your little fingernail, so their physical properties may be difficult to see without a hand lens or microscope. Further complications arise because most rock-forming mineral groups have variable compositions attributable to ionic substitution in the crystal structure. As a result, color, hardness, and other physical properties may be variable.

It is important for you to become familiar with the general characteristics of each of the major rock-forming mineral groups (feldspars, quartz, micas, olivines, pyroxenes, amphiboles, clays, calcite, dolomite, halite, and gypsum) and to know something about their physical properties, their mode of origin, the environment in which they form, and their genetic significance. Some of the characteristics of these, as well as other important but less common minerals, are listed in Table 3.3. You will find the following summary of each mineral group to be much more meaningful if you examine a specimen of a rock containing the mineral while you study the written description.

A careful examination of the minerals that make up granite is a good beginning. The polished surface of granite (Figure 3.20) shows that the rock is composed of myriad mineral grains of different sizes, shapes, and colors. Although the minerals interlock to form a tight, coherent mass, each has distinguishing properties.

Felsic Silicate Minerals

One large group of silicate minerals includes the major constituents of continental crust: feldspars and quartz. These are commonly known as **felsic minerals.** (They are sometimes called sialic because they are rich in *silicon* and *aluminum*.) In addition to being the major constituents of continental crust, the felsic minerals also have low densities and crystallize at low temperatures in magmas.

Feldspars are the most abundant minerals in granite, a common crustal rock. The granite in Figure 3.20 consists largely of a pink, porcelainous mineral that has a rectangular form and a milky-white mineral that is somewhat smaller but similarly shaped. These are feldspars (German, "field crystals"), the most abundant minerals in Earth's crust, comprising about 50%. The feldspars have good cleavage in two directions, a porcelainous luster, and a hardness of about 6 on the Mohs hardness scale. The crystal structure involves a complex three-dimensional framework of silicate tetrahedrons (Figure 3.19). Considerable ionic substitution gives rise to two major types of feldspars: potassium feldspar (K-feldspar) and plagioclase feldspar. Potassium feldspar ($KAlSi_3O_8$) is commonly pink in granitic rocks. **Plagioclase** feldspar (shown in gray in the sketch) permits complete substitution of sodium (Na) for calcium (Ca) in the crystal structure, giving rise to a compositional range from $NaAlSi_3O_8$ to $CaAl_2Si_2O_8$. Moreover, most grains of plagioclase have distinctive, closely spaced striations on their cleavage planes. Plagioclase in granite is rich in sodium. Feldspars (with a density of 2.7 g/cm^3) are common in most igneous rocks, in many metamorphic rocks, and in some sedimentary rocks. Consequently, the continental crust has a characteristically low density (ranging from 2.6 to 2.7 g/cm^3), controlled by the shear abundance of feldspar and quartz.

Quartz forms the glassy, irregularly shaped grains in Figure 3.20. It usually grows in the spaces between the other minerals. As a result, quartz in granite typically lacks well-developed crystal faces. When quartz crystals are able to grow freely, their form is elongated, has six sides, and terminates in a point, but well-formed crystals are rarely found in rocks. In sandstone, quartz is abraded into rounded sand grains.

Quartz is abundant in all three major rock types. It has the simple composition SiO_2 and is distinguished by its hardness (7), its conchoidal fracture, and its glassy luster. Pure quartz crystals are colorless, but slight impurities produce a variety of colors. Quartz is made of silicate tetrahedrons linked together in a tight framework. All of the bonds are between Si and O; it includes no other elements. As a result, quartz is very hard, and, because all of the bonds have the same strength, it lacks cleavage. Quartz is stable both mechanically (it is very hard and lacks cleavage) and chemically (it does not react with elements at or near Earth's surface). It is therefore a difficult mineral to alter or break down once it has formed.

Micas are the tiny black, shiny grains in Figure 3.20. These distinctive minerals are potassium aluminum silicates. Micas are readily recognized by their perfect one-directional cleavage, which permits breakage into thin, elastic flakes. Mica is a complex silicate with a sheet structure, which is responsible for its perfect

TABLE 3.3	Earth's Common Minerals					
Name	Composition	Cleavage/Fracture	Color	Hardness	Density (g/cm^3)	Comments
Amphibole	$Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2$	Two at 60° and 120°	Black to green	5–6	3.2	
Bauxite	$AlO(OH)$	One perfect	White	6.5	3.4	Aluminum ore, mineral diaspore
Beryl	$Be_3Al_2Si_6O_{18}$	One poor	Green, blue, red	8	2.7	Emerald is gem variety Hexagonal prisms
Biotite	$K(Mg,Fe)_3AlSi_3O_{10}(OH)_2$	One perfect	Black to dark brown	2.5–3	3	Splits into thin sheets
Calcite	$CaCO_3$	Three perfect Rhombohedral	Colorless, white	3	2.7	Bubbles in dilute acid
Chalcopyrite	$CuFeS_2$	Fracture	Brassy, golden yellow Metallic luster	4	4.3	Copper ore
Chlorite	$(Mg,Fe)_5Al_2Si_3O_{10}(OH)_8$	One perfect	Green	2	2.5	Foliated masses
Clay	$Al_2Si_2O_5(OH)_4$	One perfect	White to brown	2	2.0–2.5	Common in soils
Corundum	Al_2O_3	Fracture	Brown or blue	9	4	Rubies and sapphires
Diamond	C	Fractures	Transparent Adamantine luster	10	3.5	Hardest mineral known
Dolomite	$CaMg(CO_3)_2$	Three perfect	Transparent to white	3.5–4	2.8	Bubbles in acid when powdered
Fluorite	CaF_2	Perfect	Transparent, green, purple, yellow	4	3.2	Fluorine ore
Galena	PbS	Three perfect Cubic	Black to silver Metallic luster	2.5	7.6	Lead ore
Garnet	$Ca_3Al_2Si_3O_{12}$	Conchoidal fracture	Red to brown Vitreous luster	6.5–7	3.6	
Graphite	C	One perfect	Black	1–2	2.1	Compare with diamond
Gypsum	$CaSO_4 \cdot 2H_2O$	One perfect Two good	Transparent to white	2	2.3	Used in plasterboard
Halite	$NaCl$	Three perfect Cubic	Transparent to white	2.5	2.2	Table salt
Hematite	Fe_2O_3	None	Red to silvery gray Metallic or earthy	6	5.3	Iron ore
K-feldspar	$KAlSi_3O_8$	Two at right angles	White to gray or pink	6	2.6	
Kyanite	Al_2SiO_5	One perfect One poor	White to light blue	5–7	3.6	Long-bladed aggregates
Magnetite	Fe_3O_4	Conchoidal Irregular	Black Metallic luster	6	5.2	Magnetic
Muscovite	$KAl_3Si_3O_{10}(OH)_2$	One perfect	Colorless to light brown	2–2.5	2.8	Splits into translucent sheets
Olivine	$(Mg,Fe)_2SiO_4$	Conchoidal	Green to brown	6.5	3.4	Gem peridot
Plagioclase	$NaAlSi_3O_8$ $CaAl_2Si_2O_8$	Two at right angles	White to gray	6	2.7	Striations on cleavage planes Most common mineral at surface
Pyrite	FeS_2	Uneven fractures	Brassy to golden yellow	6.5	5	Fool's gold; well-formed cubes common
Pyroxene	$(Mg,Fe)SiO_3$	Two at about 90°	Green to dark brown or black	6	3.3	
Quartz	SiO_2	Conchoidal fracture	Colorless, also gray, purple, other	7	2.7	Six-sided elongate crystals
Serpentine	$Mg_6Si_4O_{10}(OH)_8$	Splintery fracture Asbestos fibrous	Green to brown Silky or waxy luster	2.5	2.5	
Sillimanite	Al_2SiO_5	One perfect	Colorless to white	6–7	3.2	Long, slender crystals
Staurolite	$Fe_2Al_9Si_4O_{22}(OH)_2$	One poor	Brown to red	7	3.8	
Talc	$Mg_3Si_4O_{10}(OH)_2$	One perfect	White to light green	1	2.8	Soft, soapy masses
Zeolite	Complex hydrous silicates	One perfect	Colorless to light green	4–5	2.2	Earthy, but may form radiating crystals in cavities

(A) A polished surface of a granite, shown at actual size, displays mineral grains of different sizes, shapes, and colors.
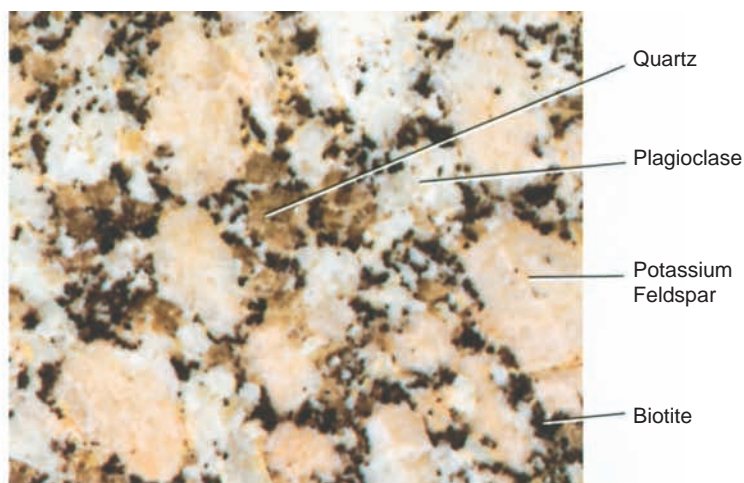
(B) An exploded diagram of (A) shows the relative size and the shape of individual mineral grains.
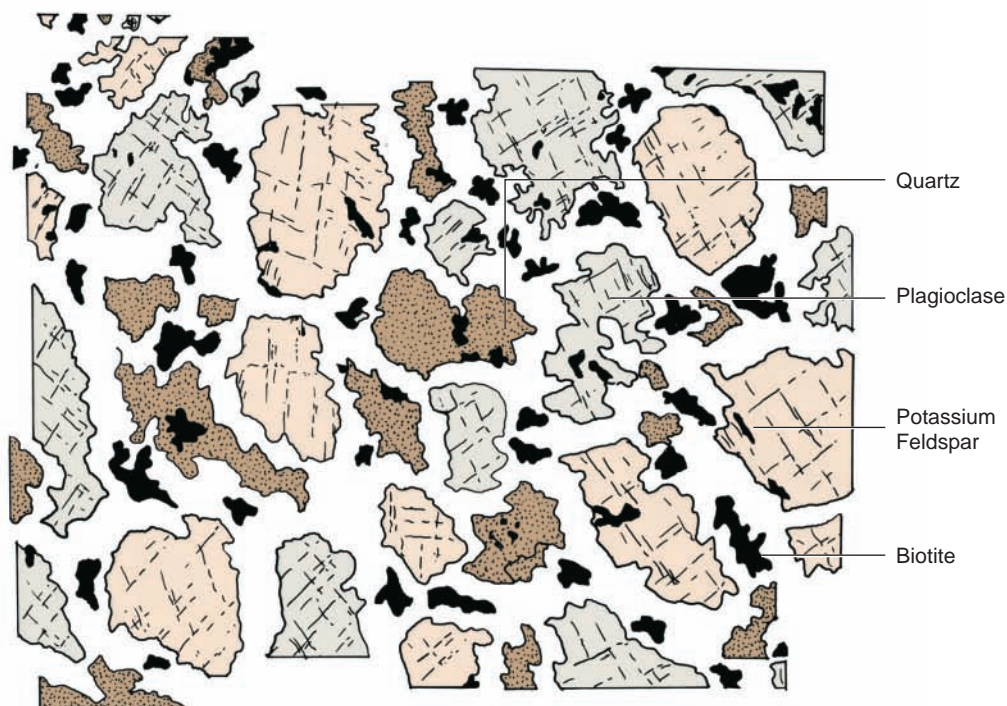
FIGURE 3.20   Mineral grains in a granite, a common rock in continental crust, form a tight, interlocking texture because each mineral is forced to compete for space as it grows. The most common minerals in granite are the felsic minerals: quartz, plagioclase feldspar, and potassium feldspar.

cleavage. Two common varieties occur in rocks: **muscovite** $[KAl_3Si_3O_{10}(OH)_2]$, which is white or colorless and is found along with felsic minerals, and **biotite** $[K(Mg, Fe)_3Si_3AlO_{10}(OH)_2]$, a black mica, rich in iron and magnesium that belongs to the category of mafic minerals discussed below. Both types of mica contain water in the form of hydroxyl ions (OH^-). The densities of these minerals are also distinctive, with biotite (about 3 g/cm^3) denser than muscovite (about 2.8 g/cm^3). Mica is abundant in granites and in many metamorphic rocks and is also a significant constituent of many sedimentary rocks.

Mafic Silicate Minerals

Another category of silicate minerals is the **mafic minerals,** so named because they contain much *magnesium* and *iron*. These minerals contrast with felsic minerals and generally range from dark green to black and have high densities. Biotite is classified in this general group, together with the olivine, pyroxene, and amphibole.
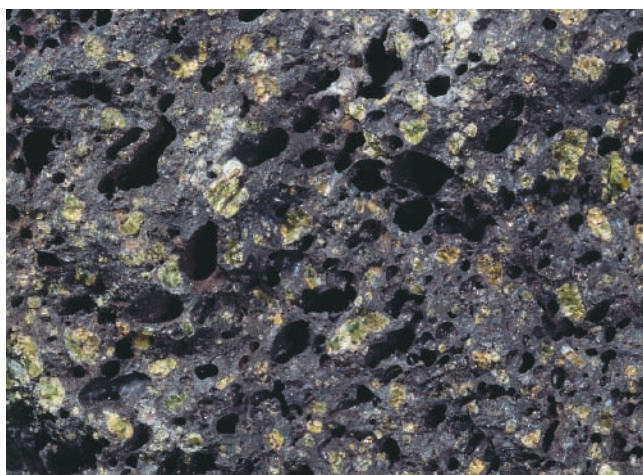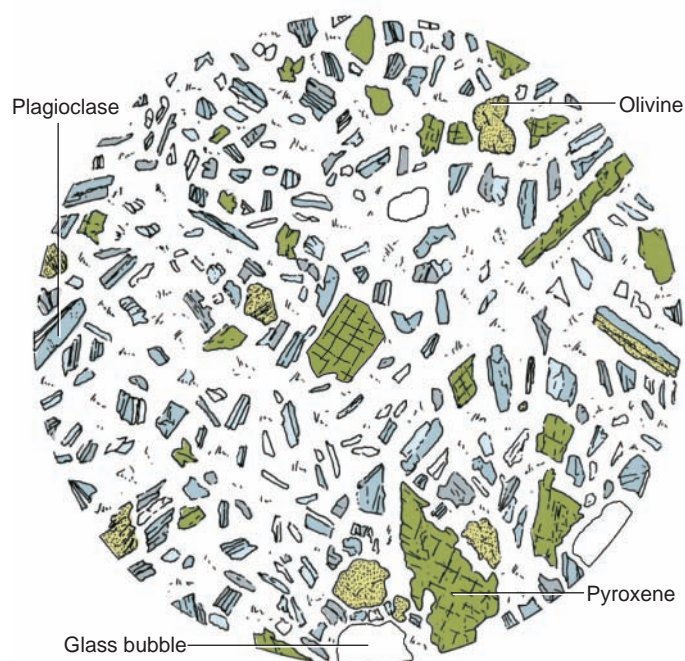
(A) In a hand specimen, only a few large grains of green olivine can be seen. The dark spots are gas bubbles frozen into the once molten rock.
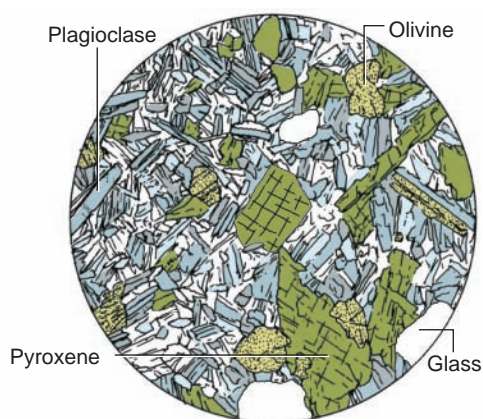
(B) Viewed through a microscope, the mineral grains form an interlocking texture. Plagioclase feldspar crystals typically form small lathlike grains between the mafic minerals.

(C) An exploded diagram of (B) shows the size and shape of individual mineral grains.

FIGURE 3.21   **Mineral grains in basalt** are microscopic and are dominated by mafic minerals. Basalt is a mafic volcanic rock common in the oceanic crust.

In granite, biotite is common, but the other mafic minerals are rare or absent. The mafic minerals are common, however, in Earth's mantle and in oceanic crust. They generally crystallize at higher temperatures and have higher densities than felsic minerals. Let us examine basalt, a common mafic volcanic rock, to see what these minerals are like (Figure 3.21).

Olivine is the only mineral clearly visible in the hand specimen in Figure 3.21; it is a green, glassy mineral. Olivine is a silicate in which iron and magnesium substitute freely in the crystal structure. The composition is expressed as $(Mg,Fe)_2SiO_4$. Olivine is composed of isolated Si-O tetrahedrons linked together by magnesium or iron ions (Figure 3.19). This hard mineral is characterized by an olive-green color (if magnesium is abundant) and a glassy luster. In rocks, it rarely forms crystals larger than a millimeter in diameter. Like most mafic minerals, olivine has a relatively high density (about 3.3 g/cm^3) and typically forms at high temperatures. It is probably a major constituent of the upper mantle. At depths of about 400 km in the mantle, olivine is no longer stable and recrystallizes to form an even denser mineral with the same elemental composition.

Pyroxenes are high-temperature minerals also found in many mafic rocks in the crust and mantle. In Figure 3.21, pyroxene occurs as microscopic crystals, but some basalts contain larger grains of this mineral, which typically range from dark

FIGURE 3.22   Amphibole crystals were among the first to crystallize in this "granitic" rock and therefore have well-developed crystal faces. The largest grain is about 3 cm long.

green to black. Their internal structure consists of single chains of linked Si-O tetrahedrons (Figure 3.19). Pyroxene crystals commonly have two directions of cleavage that intersect at right angles.

Amphiboles (Figure 3.22) have much in common with the pyroxenes. Their chemical compositions are similar, except that amphiboles contain hydroxyl ions (OH$^-$) and pyroxenes do not. The minerals also differ in structure. The internal structure consists of double chains of silicon-oxygen tetrahedrons (Figure 3.19). The amphiboles produce elongate crystals that cleave perfectly in two planes, which are not at right angles. Amphibole ranges from green to black. This mineral is common in many igneous and metamorphic rocks. Hornblende [$NaCa(Mg,Fe)_5AlSi_7O_{22}(OH)_2$] is the most common variety of amphibole. The density of a typical amphibole is about 3.2 g/cm^3.

A dangerous form of amphibole is asbestos, once used widely to make fireproof fabrics, tiles, and as insulation in buildings. Miners working in old dusty mines became sick as small cleavage fragments of a specific type of this mineral became lodged in their lungs, especially in conjunction with cigarette smoking. The incidence of this noncancerous lung disease in modern mines with dust controls is much lower. Fortunately, most asbestos used in construction consists of an entirely different mineral, and the hazard to people is much less than commonly supposed.

Clay Minerals

The **clay minerals** form another important group of silicate minerals. They are a major part of the soil and are thus encountered more frequently in everyday experience than many other minerals. Clay minerals form at Earth's surface, where air and water react with various silicate minerals, breaking them down to form clay and other products. Like the micas, the clay minerals are sheet silicates (Figure 3.19), but their crystals are usually microscopic and are most easily detected with an electron microscope (Figure 3.15). More than a dozen clay minerals can be distinguished on the basis of their crystal structures and variations in composition. A common clay mineral, kaolinite, has the formula $Al_4Si_4O_{10}(OH)_8$ and a low density of about 2.6 g/cm^3.

Nonsilicate Minerals

Some important rock-forming minerals are not silicate minerals. Most of these minerals are carbonates or sulfates and typically form at low temperatures and pressures near Earth's surface.

Calcite is composed of calcium carbonate ($CaCO_3$), the principal mineral in limestone. It can precipitate directly from seawater or is removed from seawater by organisms as they use it to make their shells. Calcite is dissolved by groundwater and reprecipitated as new crystals in caves and fractures in rock. It is usually transparent or white, but the aggregates of calcite crystals that form limestone contain various impurities that give them gray or brown hues. Calcite is common at Earth's surface and is easy to identify. It is soft enough (hardness of 3) to scratch with a knife, and it effervesces in dilute hydrochloric acid. It has perfect cleavage in three planes, which are not at right angles, so that cleaved fragments form rhombohedra (see Figure 3.9). Besides being the major constituent of limestone, calcite is the major mineral in the metamorphic rock marble. Calcite has a density of about 2.7 g/cm^3.

Dolomite is a carbonate of calcium and magnesium [$CaMg(CO_3)_2$]. Large crystals form rhombohedra, but most dolomite occurs as granular masses of small crystals. Dolomite is widespread in sedimentary rocks, forming when calcite reacts with solutions of magnesium carbonate in seawater or groundwater. Dolomite can be distinguished from calcite because it effervesces in dilute hydrochloric acid only if it is in powdered form. Dolomite has a density of nearly 2.9 g/cm^3.

Halite and **gypsum** are the two most common minerals formed by evaporation of seawater or saline lake water. Halite, common salt (NaCl), is easily identified by its taste. It also has one of the simplest of all crystal structures; the sodium and chloride ions form a cubical array. Most physical properties of halite are related to this structure. Halite crystals cleave in three planes, at right angles, to form cubic or rectangular fragments (Figure 3.9). Salt, of course, is very soluble and readily dissolves in water.

Gypsum is composed of calcium sulfate and water ($CaSO_4 \cdot 2H_2O$). It forms crystals that are generally colorless, with a glassy or silky luster. It is a very soft mineral and can be scratched easily with a fingernail. It cleaves perfectly in one plane to form thin, nonelastic plates (Figure 3.10). See the GeoLogic discussion at the end of the chapter for more information about the internal structure of gypsum. Gypsum occurs as single crystals, as aggregates of crystals in compact masses (alabaster), and as a fibrous form (satin spar).

Oxide minerals lack silicon as well and include several economically important iron oxides, such as magnetite and hematite (Table 3.3). Magnetite is particularly interesting because it is one of only a very few minerals that are naturally magnetic.

A wide variety of other minerals have been identified, including silicates, carbonates, oxides, sulfides, and sulfates. There are literally thousands of naturally formed minerals; some seem rare and exotic because of their color, crystal form, and hardness, and others seem more mundane because they occur as minor constituents in common rocks. Some we consider precious, such as gold, silver, diamonds, and rubies; others are important in high technology. In addition to providing documents of Earth's history, minerals are at the foundation of all human societies—from pre-Paleolithic times, in which minerals were used for tools, to modern technological societies that require vast amounts of metals and construction materials.
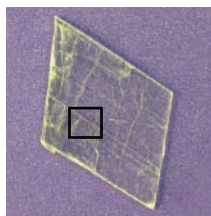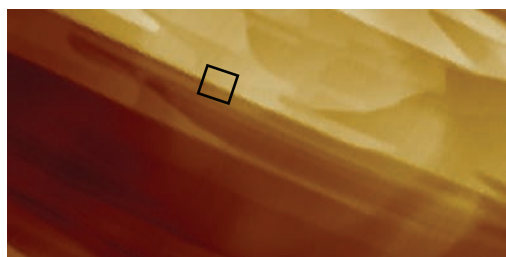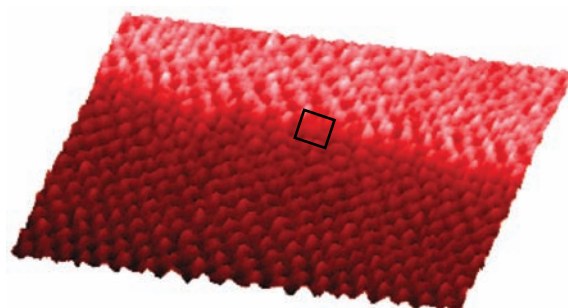
Gypsum ($CaSO_4 \cdot 2H_2O$) (1 cm across). Cleavage planes look like topographic steps
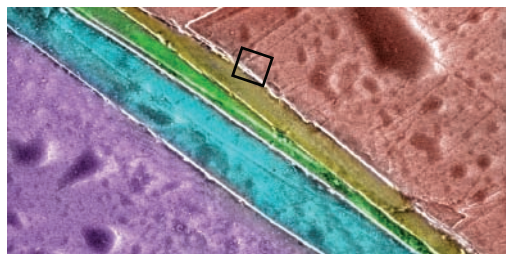

Gypsum seen with scanning electron microscope (150 microns across). Cleavage planes still visible as plateaus separating different cleavage sheets.

Gypsum seen with atomic force microscope (10 microns across). Each cleavage sheet consists of even thinner layers.

Gypsum seen with scanning tunneling microscope (15 nanometers across). Individual sulfate ions seen as hills in precise geometric arrangement. Lighter area is an atom high "plateau," magnified from the sheets seen in the cleavage planes.
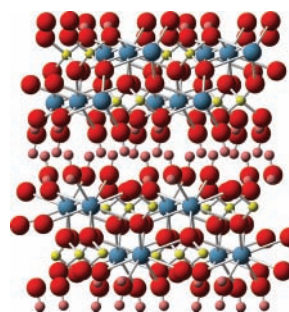
Atomic model of the internal sturctue of gypsum. Calcium ions (blue) are linked to sulfate groups (sulfur yellow and oxygen red). Sulfur-oxygen bonds are strong and covalent. Cleavage planes are created by weak bonds between these tightly bonded sheets. The layers are bound together by weak hydrogen bonds. (Hydrogens are small and pink.)

(Images courtesy of Dirk Bosbach and Barry Bickmore)

Seeing is believing—a phrase we often use to discount the unseeable. But how can we understand the internal structure of minerals at the atomic scale where distances are measured in nanometers (10^{-9} m)? These images take you on a tour through inner space, from the surface of a mineral into its deep interior.

Observations

1. At the lowest magnification with an optical microscope, you can see the nature of a cleavage plane in gypsum—a smooth lustrous break.
2. If we zoom in closer with a scanning electron microscope, you can see that the cleavage plane is not quite as smooth as it first looked, but you can still see broad flat plateaus.
3. Zooming in closer with an atomic force microscope, we can see that the planar structure is preserved at the micron scale (10^{-6} m). You can see that the mineral grew as a series of layers controlled by its internal structure. Each "layer" has a relatively smooth surface but is only a molecule thick.
4. At even higher resolution, the atomic force microscope

shows the "smooth surface" is a series of humps and swales—individual groups of atoms—packed into a precise geometric network. The "step" shown in brighter colors is one atomic layer thick and lies on top of other similar layers below it.

Interpretations

The last step of the journey is not a real image, but rather an interpretive model that has been constructed from the information gleaned by studying the other images and from X-ray diffractometry. Each group of atoms is bound together by a strong electrical charge emanating from a cloud of electrons. This electron cloud gives the atomic groups their shapes in the images. Finally, the images and measurements show us that the physical properties of a mineral are controlled by its internal atomic structure. For example, strong bonds form hard minerals and where weak bonds are aligned like those between the hydrogen ions shown in the model, the mineral cleaves easily in that direction.

KEY TERMS

amorphous solid (p. 57)	crystal form (p. 64)	isotope (p. 56)	plagioclase (p. 72)
amphibole (p.76)	crystallization (p. 67)	liquid (p. 57)	polymorphism (p. 62)
atom (p. 54)	crystal structure (p. 57)	luster (p. 65)	proton (p. 55)
atomic mass (p. 55)	density (p. 65)	mafic mineral (p. 74)	pyroxene (p. 75)
atomic number (p. 55)	dolomite (p. 77)	magnetism (p. 66)	quartz (p. 72)
biotite (p. 74)	electron (p. 55)	melt (p. 68)	recrystallization (p. 69)
calcite (p. 77)	feldspar (p. 72)	metallic bond (p. 57)	silicates (p. 70)
clay mineral (p. 76)	felsic minerals (p. 72)	metastable (p. 66)	silicon-oxygen tetrahedron
cleavage (p. 64)	gas (p. 58)	mica (p. 72)	(p. 70)
color (p. 65)	gypsum (p. 77)	mineral (p. 59)	solid (p. 57)
compound (p. 56)	halite (p. 77)	muscovite (p. 74)	stability range (p. 66)
conchoidal fracture (p. 64)	hardness (p. 65)	neutron (p. 55)	stable (p. 66)
covalent bond (p. 56)	ion (p. 56)	nucleus (p. 54)	streak (p. 66)
crystal (p. 61)	ionic bond (p. 56)	olivine (p. 75)	X-ray diffraction (p. 60)
crystal faces (p. 64)	ionic substitution (p. 62)	oxide mineral (p. 77)	

REVIEW QUESTIONS

1. Contrast atoms, ions, and isotopes.
2. Give a brief but adequate definition of a mineral.
3. Explain the meaning of "the internal structure of a mineral."
4. Why does a mineral have a definite chemical composition?
5. What other common element might substitute for Ca in a plagioclase feldspar? Why?
6. How do geologists identify minerals too small to be seen in a hand specimen?
7. Briefly explain how minerals grow and are destroyed.
8. Explain the origin of cleavage in minerals.
9. Describe the silicon-oxygen tetrahedron. Why is it important in the study of minerals?
10. Discuss the implications of a mineral's limited stability range for the kinds of minerals found at progressively greater depths in Earth's mantle.
11. Why is color of little use in identifying minerals? What are some better diagnostic properties?
12. What are silicate minerals? List the silicate minerals that are most abundant in rocks.
13. Why are feldspars so abundant in Earth's crust?
14. Construct a table listing the distinguishing characteristics of quartz, feldspar, biotite, amphibole, pyroxene, mica, and clay.
15. Study Figure 3.20, and explain why most of the mineral grains in a granite have an irregular shape even though they still have an orderly atomic structure.
16. What is the difference between a mineral and a rock?

ADDITIONAL READINGS

Deer, W. A., R. A. Howie, and J. Zussman. 1992. *An Introduction to the Rock-Forming Minerals,* 2nd ed. New York: Wiley.

Klein, C. 2002. *Manual of Mineral Science (after J. D. Dana),* 22nd ed. New York: Wiley.

Mackenzie, W. S., and A. E. Adams. 1994. *A Color Atlas of Rocks and Minerals in Thin Section.* New York: Halstead Press.

Nesse, W. D. 2000. *Introduction to Mineralogy.* New York: Oxford University Press.

Perkins, D. 2001. *Mineralogy*, 2nd ed. Upper Saddle River, N.J.: Prentice Hall.

Riciutti, E. R. 1998. *National Audubon First Field Guide to Rocks and Minerals.* New York: Scholastic.

MULTIMEDIA TOOLS

Earth's Dynamic Systems Website
The Companion Website at **www.prenhall.com/hamblin** provides you with an on-line study guide and additional resources for each chapter, including:

- On-line Quizzes (Chapter Review, Visualizing Geology, Quick Review, Vocabulary Flash Cards) with instant feedback
- Quantitative Problems
- Critical Thinking Exercises
- Web Resources

Earth's Dynamic Systems CD
Examine the CD that came with your text. It is designed to help you visualize and thus understand the concepts in this chapter. It includes:

- Animations showing the three-dimensional atomic structures of common silicate and nonsilicate minerals
- Video clips of mineral growth
- A direct link to the Companion Website