

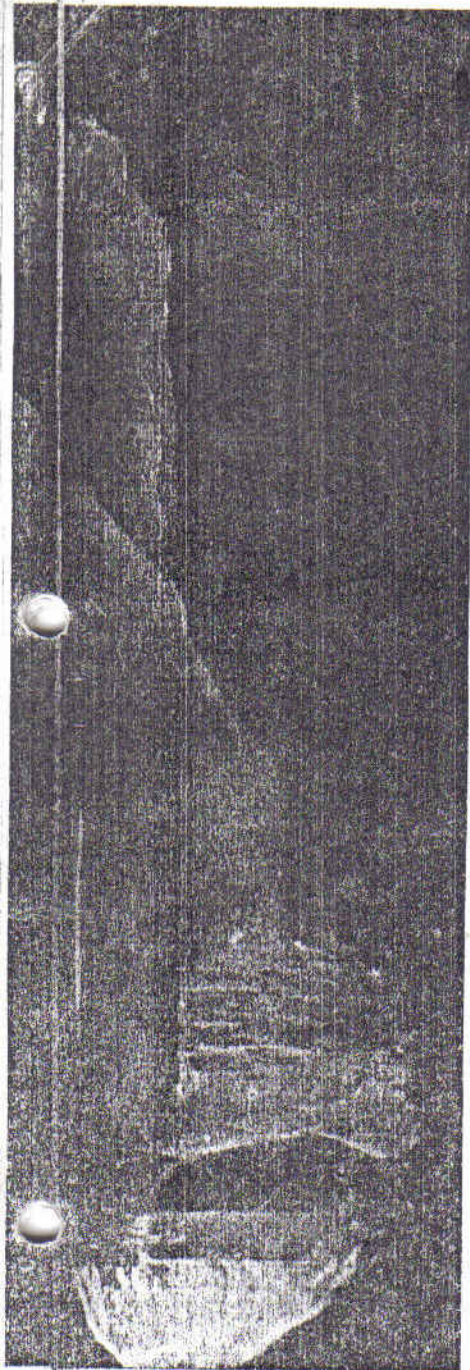
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Minerals

LECTURE 6L41C

Fri 22/08/2002

Multicolor tourmaline crystals known as the "Rabbit Ears," grown on quartz crystals (right). Specimen collected at the Himalaya Mine, San Diego County, California. Tourmaline from the permanent collections of the Houston Museum of Natural Science.



The solid part of the Earth is composed of special types of compounds called minerals, which range from spectacularly beautiful rubies and emeralds to dull, earthy “rust.” In this chapter we will discuss what makes minerals unique, examine the processes by which they form, and introduce those that are most common in the Earth’s crust. We also will show how geologists use minerals to interpret Earth processes.

WHAT IS A MINERAL?

To some, anything that is neither animal nor vegetable must be a mineral. To others, minerals are related to vitamins and must be part of our diet if we are to stay healthy. To a *geologist*, however:

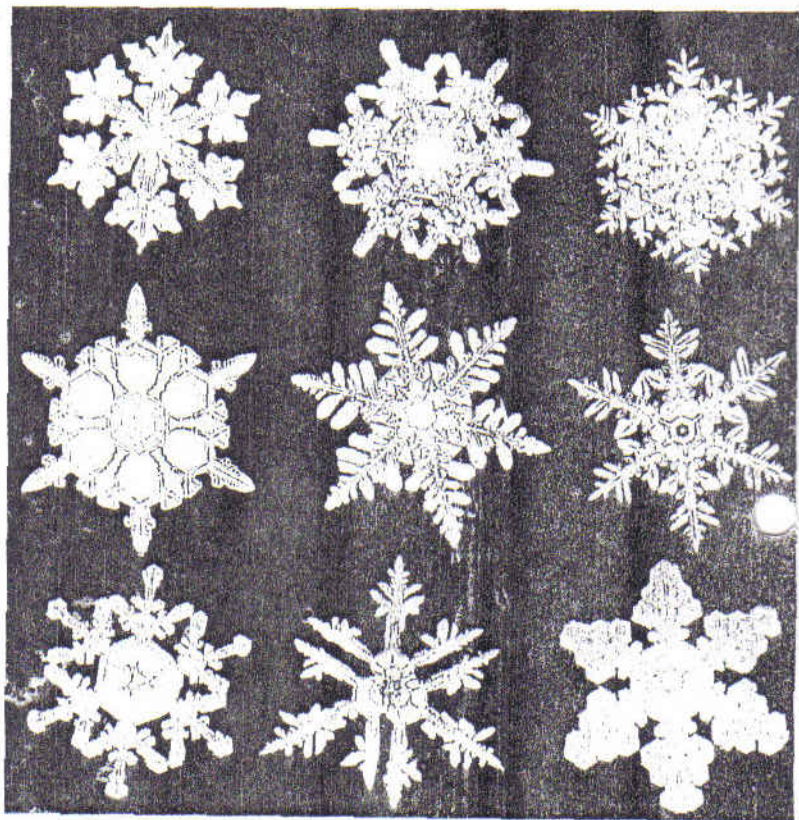
A **mineral** is a naturally occurring, inorganic solid with an ordered internal arrangement of atoms or ions and a specific chemical composition.

By definition, all minerals are solids, so neither water nor gases can be minerals. Some minerals are native elements such as gold, sulfur, and diamond (carbon). Some are simple compounds, such as halite (NaCl). Others have complex formulas that are longer than a line in this book.

All minerals form by chemical reactions and are held together by one of the bond types, or some combination of bond types, described in Chapter 2. When we apply the definition of a mineral to Earth materials, some gems and “mineral resources” prove not to be minerals at all. Amber (hardened sap of ancient trees), pearls (secretions of oysters), and ivory (the tusks of elephants and walruses) are not minerals because they are always the result of organic processes. Oil and natural gas are not solids and should not be called mineral resources. Opal is a naturally occurring inorganic solid, but it is only a *mineraloid* because it lacks the rigidly ordered internal structure and specific chemical composition required of minerals.

FIGURE 3-1

The complex crystal shapes of snowflakes. Note the characteristic sixfold symmetry. Photo by American Museum of Natural History.



Several common materials do satisfy the definition. Salt is a mineral, a natural inorganic solid with ordered internal arrangement of its specific ions, sodium and chlorine. A snowflake is a mineral, forming naturally by inorganic processes, with a constant chemical composition, and with its hydrogen and oxygen atoms in fixed positions within a geometrically ordered structure. It is this rigid structure that yields the complex but always symmetrical shapes of snowflakes (Figure 3-1). Their *ordered internal structure* is what sets minerals apart from other types of compounds, and we shall discuss that structure in detail in the following sections.

WHAT'S IN A NAME?

Minerals, as compounds, could be identified by their chemical formulas, but they are normally called by names, either scientific or common. For example, sodium chloride (NaCl) is the mineral halite, commonly called "rock salt." Silicon dioxide (SiO_2) is the mineral quartz.

Mineral names come from many sources. Some are named for where they were discovered: labradorite was named for Labrador in Canada. Other names describe a mineral's appearance or physical properties: azurite is deep blue (azure); orthoclase is named from the Greek words *orthos* (meaning straight) and *klastos* (meaning broken) because, when struck, it forms fragments bounded by 90° angles. Minerals also are named after famous people: sillimanite after Benjamin Silliman, an early American mineralogist, and armalcolite after the first *Apollo* astronaut crew to visit the Moon (Armstrong, Aldrin, and Collins).

THE INTERNAL STRUCTURE OF MINERALS

Under favorable growth conditions, minerals form regular geometric shapes called **crystals** (Figure 3-2). Some are simple shapes with only a few faces, but others, such as snowflakes, are more complex. In 1669, after years of studying crystals, the Danish scientist Nicolaus Steno proposed that all minerals possess an ordered internal structure. He found that each of the hundreds of quartz crystals in his collec-

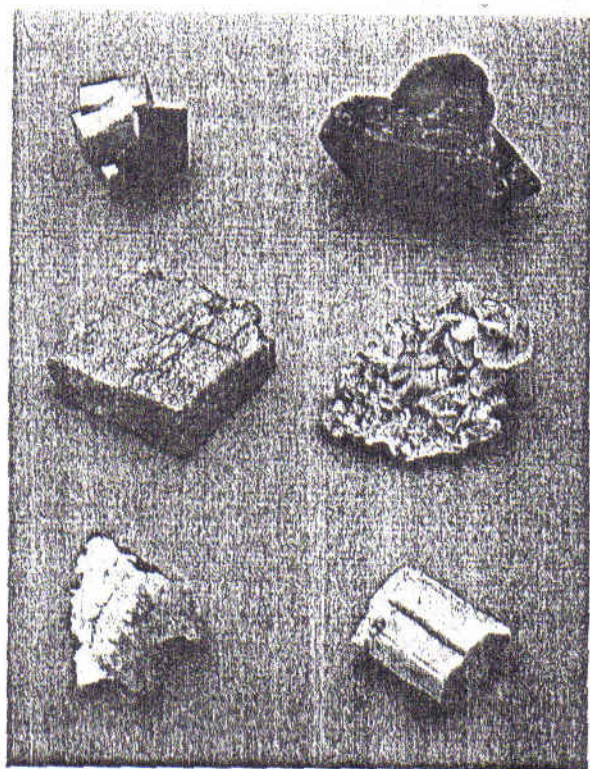


FIGURE 3-2

The shapes of mineral crystals vary widely. Clockwise from upper left, examples shown are cubes of pyrite, six-sided prisms and pyramids of smoky quartz, tabular crystals of wulfenite, prisms of beryl, stubby prisms of apophyllite, and rhomb of rhodochrosite. Macmillan Publishing/Geoscience Resources photo.

tion had the same symmetrical arrangement of crystal faces and had identical angles between faces, regardless of their size, color, or degree of perfection. He also observed that crystals of galena were identical to one another, but had a different kind of symmetry than quartz. Steno reasoned that this could happen only if a very small, basic unit were arranged in the same way within every quartz crystal, and similarly within every galena crystal, although a different arrangement. Steno knew nothing of atoms, of course—they were discovered long after his death.

Two hundred fifty years later we learned that Steno was right. In 1912 German physicist Max von Laue irradiated minerals with X-rays, themselves a new discovery. He correctly deduced that the resulting geometric patterns (Figure 3-3) were caused by X-rays interacting with ordered planes of atoms.

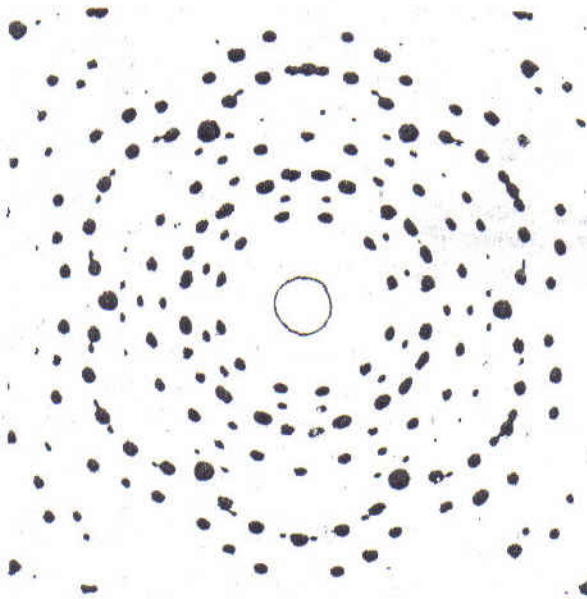
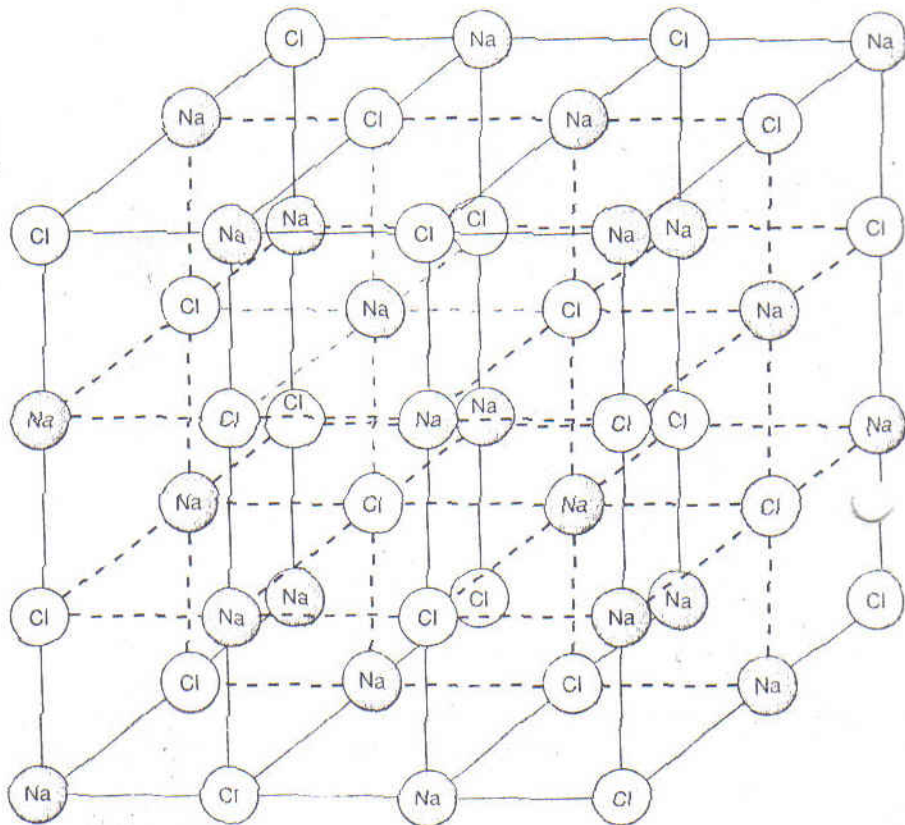


FIGURE 3-3

This regular, geometric pattern results when X-rays pass through the ordered atomic structure of a mineral. Photo by Eastman Kodak company.

FIGURE 3-4

The internal atomic structure of halite or sodium chloride, NaCl . This ordered array of alternating Na^+ and Cl^- ions was interpreted in 1914 by W. H. Bragg and W. L. Bragg from X-ray images like Figure 3-3.



Within two years, the British father-and-son team of W. H. Bragg and W. L. Bragg successfully determined the geometrically perfect internal arrangement of sodium and chlorine ions in the mineral *halite* (Figure 3-4) from its X-ray pattern.

Today, mineralogists using sophisticated X-ray instruments have learned the structures of most common minerals. The ordered structure of minerals is called a **crystalline structure** because it was discovered first in well-formed crystals, but X-rays show that it is present in every sample of a mineral. Some *mineral specimens* grow too rapidly or in an environment too crowded with other specimens to form crystals, but even these apparently shapeless masses possess the internal structure characteristic of the specific mineral. Each mineral's structure is a unique three-dimensional arrangement of particular atoms or ions and can be used like fingerprints to help in *mineral identification*.

Why Do Minerals Have Crystalline Structures?

Atoms and molecules in gases and liquids can move freely. For example, once two oxygen atoms are joined by strong covalent bonds to form an O_2 molecule, only weak van der Waals' forces hold the molecule to other molecules. This allows freedom of movement, for each oxygen atom is bonded strongly to only one other atom. In the mineral kingdom, however, each atom or ion is firmly bonded to *several* others, and this locks them in place. In the mineral halite (NaCl), for example, each sodium ion is bonded to six chlorine ions, and each chlorine to six sodiums (Figure 3-4). The reason for this arrangement lies in the sizes and charges of the ions.

Ionic size and charge. The halite structure demonstrates the role of *ionic size* in controlling a mineral's internal arrangement of atoms or ions. Imagine an environment in which millions of sodium atoms lose electrons to millions of chlorine atoms, forming sodium cations (+) and chlorine anions (-). Each sodium ion then attracts as many chlorine ions as can fit around it—*not just the one involved in the electron transfer*. Similarly, each chlorine ion attracts as many sodium ions as can surround it.

The number of anions that can fit around a cation is called the **coordination number** and depends only on the relative sizes of the ions involved. Each anion barely touches the surface of the cation, and the anions repel one another so that none actually come into contact. In halite, six Cl^- ions surround every Na^+ ion; they lie at the corners of an eight-sided figure, with the sodium at the center (Figure 3-5). This geometrically perfect arrangement is found in every sample of halite and results in the same basic crystal shape every time halite grows.

In halite, the six chlorine ions surrounding sodium act to neutralize its +1 charge. Each Cl^- uses one-sixth of its total -1 charge to do so, leaving it with five-sixths of a negative charge. Every Cl^- ion must then find five more sodium ions with which to bond in order to reach electrical neutrality itself. Like all compounds, NaCl is electrically neutral. Its formula shows that equal numbers of Na^+ and Cl^- ions are present, but the *sizes and charges* of these ions dictate the physical geometry by which the ion-by-ion neutrality is achieved.

The manner in which each mineral manages its coordination and charge balance depends on the combination of ions involved. Ions vary considerably in size, as shown in Table 3-1. Those with the greater number of electron shells are the largest and those with fewest shells are the smallest. There are many possible combinations and several kinds of coordination, including the fourfold coordination of oxygen around silicon (Figure 3-5) for the two most common ions in the Earth's crust. As a result, there are many kinds of mineral structures.

Polymorphs: The Role of Temperature and Pressure.

Diamond is the hardest mineral. It is translucent and may be colorless, yellow, blue, or green. Graphite, on the other hand, is one of the softest minerals. It is opaque, gray-black, and has a greasy feel. Although their appearances and physical properties are dramatically different, both minerals have exactly the same chemical composition: each is made entirely of carbon. Graphite and diamond are **polymorphs**, minerals that have identical compositions but different internal structures.

Diamond and graphite form under very different conditions, and their internal structures reflect this (Figure 3-6). Diamond's structure is much more compact than that of graphite because diamond

TABLE 3-1

Ionic size and coordination number.






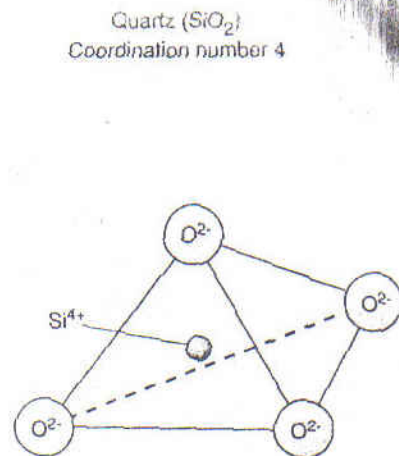
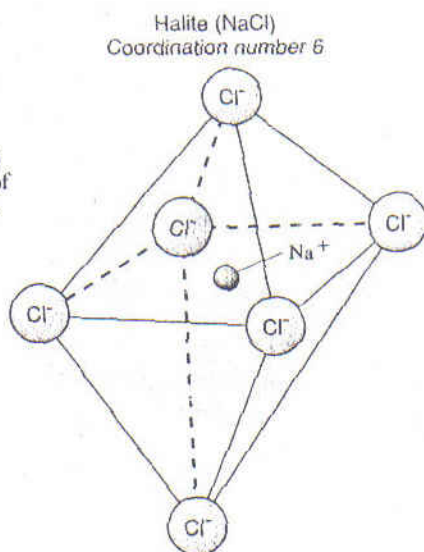
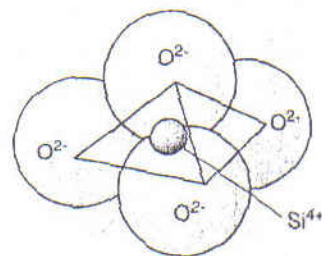
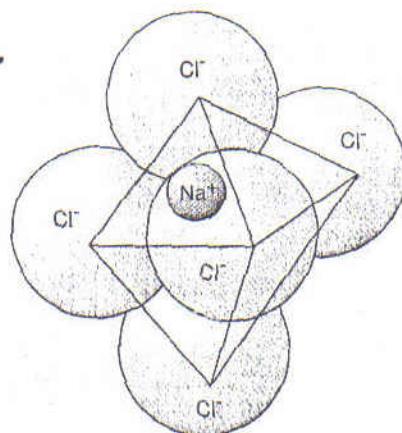
Relative ion size	Approximate ionic radius $\times 10^{-8}$ cm	Coordination with oxygen	Possible substitutions
	Oxygen 1.40		
	Silicon 0.34	4	Si^{4+} , Al^{3+}
	Magnesium 0.66	6	Al^{3+} , Mg^{2+} , Fe^{2+} , Mn^{2+} , Fe^{3+} , Cr^{3+}
	Calcium 0.99	8	Ca^{2+} , Na^+ , Sr^{2+}
	Potassium 1.33	12	Na^+ , K^+ , Rb^+

FIGURE 3-5

The coordination principle. Every ion is surrounded by as many ions of the opposite charge as can be packed around it. The number of neighbors (the *coordination number*) depends only on the relative sizes of the cations and anions involved. In addition to the examples shown, coordination numbers of 2, 3, 8, and 12 are possible.



(a) Stick models showing positions of ions



(b) Models showing relative ionic sizes

forms under extremely high pressures. Each carbon atom in diamond is bonded covalently to *four* others, forming a close-knit, three-dimensional framework. In graphite, each carbon is covalently bonded to *three* others, forming a series of two-dimensional sheets. These sheets are then held together by weak van der Waals' forces. As a result of their different structures, the density of diamond (the mass per unit volume) is much greater than that of graphite: 3.5 g/cm^3 (grams per cubic centimeter) for diamond versus 2.3 g/cm^3 for graphite.

Polymorphs such as diamond and graphite are useful for determining the physical conditions (pressure, temperature) of Earth processes and the locations in the Earth where minerals form. Experiments show that the pressures needed to form diamond are greater than those found anywhere in the Earth's crust, so rocks containing diamond must have originated in the mantle and somehow moved close enough to the surface that we could mine them. Such processes are rare, so it is no wonder that diamonds are scarce and therefore precious.

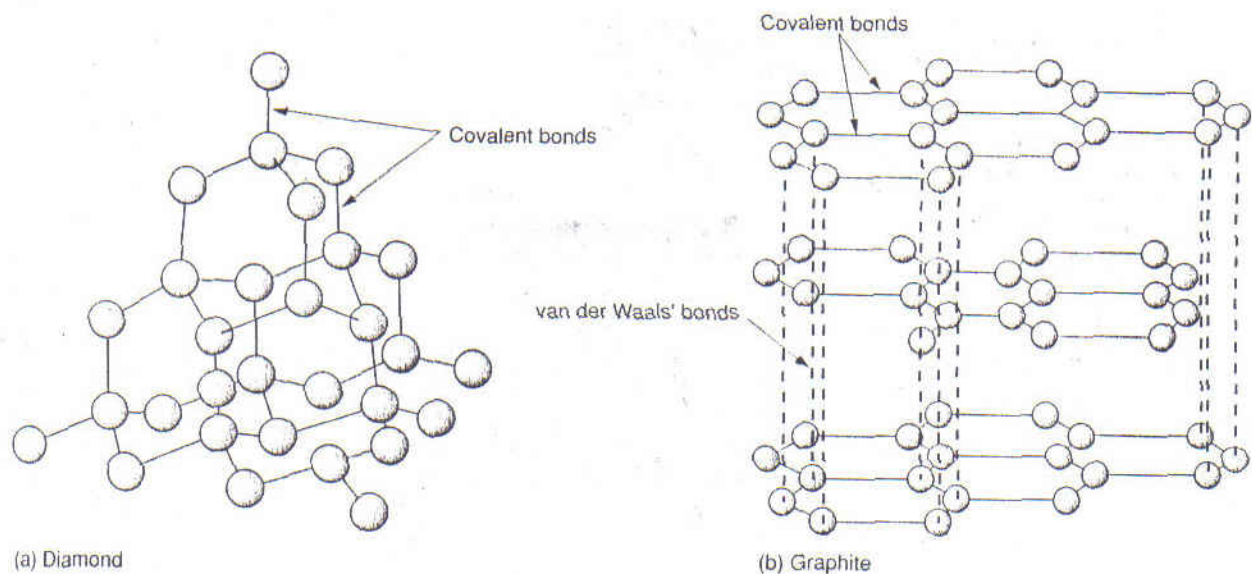


FIGURE 3-6

The internal structures of graphite and diamond. (a) In diamond, all carbon atoms are covalently bonded to one another in a dense, tightly packed structure. This accounts for diamond's tremendous hardness. (b) In graphite, the covalent bonds link carbon atoms in two-dimensional sheets; the sheets are held to one another by weak van der Waals' bonds. These bonds are so easily broken that the graphite sheets can slide against one another, making graphite a slippery lubricant.

VARIATIONS IN MINERAL COMPOSITION

Minerals such as graphite (C), quartz (SiO_2), and halite (NaCl) have fixed chemical compositions. However, nearly every mineral contains some impurities in its structure; most impurities are ions that were trapped during the mineral's growth. These impurities may bring about variations in a mineral's physical properties. Pure quartz, for example, is colorless, but various impurities turn it pink, purple, green, yellow, or other colors.

Olivine, the mineral known to jewelers as peridot, is different because it has a systematic variation in composition rather than a random inclusion of stray ions. In some mineral structures, several common ions can take the place of certain others, a process called **ionic substitution**.

Ionic Substitution

The olivine group of minerals, represented by the formula $(\text{Fe}, \text{Mg})_2\text{SiO}_4$, is a common example of

ionic substitution. The parentheses indicate ions that can substitute for one another (Mg and Fe) and indicate that the two can be present in any proportion. Pure Mg_2SiO_4 can exist as the mineral forsterite, or pure Fe_2SiO_4 as the mineral fayalite, and so can minerals with any composition intermediate between the two, such as FeMgSiO_4 .

We saw earlier that the size and charge of ions determine mineral structure. Oxygen is the only anion in olivines and must surround all the cations present. The silicon ion, Si^{4+} , is small and is coordinated by only four oxygens, but both iron, Fe^{2+} , and magnesium, Mg^{2+} , ions are of the appropriate size to be surrounded by six oxygens. They also have the same +2 charge. As a result, either can fit into the same place in the olivine structure. Indeed, iron and magnesium are interchangeable in many minerals and are said to substitute for one another. In general, *two ions can substitute easily for one another if they have the same charge and size.* (Note that neither Fe^{2+} nor Mg^{2+} can substitute for silicon—they are too large to fit into its structural position.)

The olivine group is an example of a **solid-solution series**, a family of minerals that can have any composition between two extremes, called **end members**, and still maintain the same structure. Forsterite and fayalite are said to be end members of the *olivine solid-solution series*.

Coupled Ionic Substitution. Even if they do not have the same charge, ions can substitute for one another as long as they are the same size and some way can be found to maintain *electric neutrality*. The plagioclase feldspar solid-solution series, the most abundant mineral group in the Earth's crust, is an example of this kind of chemical variation.

The end members of the plagioclase feldspar series are albite, $\text{NaAlSi}_3\text{O}_8$, and anorthite, $\text{CaAl}_2\text{Si}_2\text{O}_8$. Ca^{2+} and Na^+ are so similar in size that they commonly substitute for each other in minerals, even though their charges are different. When a calcium ion substitutes for sodium in albite, an extra positive charge is added, but this problem is solved by simultaneously substituting an aluminum ion, Al^{3+} , for a silicon ion, Si^{4+} . Electric neutrality is maintained because the combined charges of $\text{Ca}^{2+} + \text{Al}^{3+}$ —five—equal the charges of the $\text{Na}^+ + \text{Si}^{4+}$ that they replace. This type of substitution is called **coupled ionic substitution** because one change requires (or is coupled to) the other.

IDENTIFICATION OF MINERALS

A mineral's unique combination of chemical composition and internal structure produces a diagnostic set of physical properties. These properties allow us to identify an unfamiliar mineral without using the sophisticated or expensive equipment needed for chemical or structural analyses. In most instances, the properties are due to the presence of a specific atom or ion in the mineral or to a particular type of bonding.

The physical properties useful in mineral identification include crystal shape, color, luster, hardness, the manner in which a specimen breaks, and specific gravity. Other helpful attributes are less familiar: streak (the color of a mineral's powder), magnetism, malleability, and ductility.

Crystal Shape (Habit)

We have seen that a mineral's internal structure results in a regular crystal form, if growth conditions permit. This crystal form is called a mineral's **habit** and is useful in identifying many minerals (Figure 3-2). Garnets, for example, form 12-sided or 24-sided crystals, but never the cubes favored by halite or the elongate hexagonal (six-sided) crystals of quartz. When unfavorable growth conditions pre-

TABLE 3-2
Mohs scale for mineral hardness.

Mohs Scale		Common Testing Materials	
Hardness Number ¹	Standard Mineral	Material	Hardness
10	Diamond		
9	Corundum		
8	Topaz, beryl		
7	Quartz	← Streak plate	7
		← Hard steel file	6.5
6	Orthoclase	← Window glass	5.5
5	Apatite	← Pocketknife	5.0
4	Fluorite		
3	Calcite	← Penny	3.0
		← Fingernail	2.5
2	Gypsum		
1	Talc		

¹This is a scale of *relative* hardness. It shows that quartz is harder than anything lower than 7 and softer than anything higher than 7. In terms of *absolute* hardness, corundum is twice as hard as topaz and nearly 400 times harder than talc.

vent the formation of crystal faces, other properties must be used for identification.

Hardness

Hardness is the resistance of a mineral to being scratched and indicates the bond strength in a mineral. Diamond is harder than quartz because the bonds holding diamond's carbon atoms together are stronger than the forces that bond silicon to oxygen in quartz. We can determine hardness by seeing whether a mineral can scratch (or be scratched by) standard test minerals on a reference scale called **Mohs hardness scale**. This scale, devised by German mineralogist Friedrich Mohs in 1812, remains the standard for determining mineral hardness (Table 3-2). For example, a hardness of 5.5 on the Mohs scale means that a mineral is harder than apatite but softer than orthoclase.

As shown in the table, household materials such as a steel knife or a piece of glass also can be used, once their hardnesses are calibrated against the scale. A mineral will scratch any substance softer than itself, but will be scratched by any harder substance. For example, garnets have a hardness around 7. They will scratch a glass plate (hardness 5.5–6.0) but not a ruby (a gem-quality form of corundum with a hardness of 9).

In some minerals, bonds are stronger in certain directions, and care must be used in testing hardness. Kyanite is one of the best examples of such directional hardness. A pocketknife (hardness 5) can scratch kyanite if used parallel to the long dimension of its crystal (hardness 4–5), but cannot scratch it if used parallel to the short dimension (hardness 6–7).

Cleavage

Minerals break in distinctive ways. For example, minerals of the mica family can be peeled into smooth, thin sheets (Figure 3-7). Feldspars, important constituents of many rocks such as granite, also break along smooth surfaces when struck, but they do not break into sheets. The feldspar breakage surfaces are at nearly right angles to one another (like the halite in Figure 3-8). The tendency of these minerals to break along sets of parallel planes is called **cleavage**, and the surfaces are called **cleavage planes**.

Cleavage occurs in those minerals that have planar zones of weak bonding. In graphite, for example,

cleavage occurs because weak van der Waals' bonds are easily broken. As a result, the cleavage planes are parallel to the sheets of carbon atoms shown in Figure 3-6.

To describe cleavage, it is necessary to note two things: the *number of different cleavage directions* (two in feldspars, one in micas) and the *angles between the cleavage directions*. Two minerals might each have two directions of cleavage, but the cleavage planes in one could be at right angles (as in halite), whereas the cleavage planes in the other could be nonperpendicular (as in calcite—Figure 3-8).

Fracture

Some minerals do not cleave at all because they have no planar zones of weak bonds in their structures. Such minerals break by what is called **fracture**. Fracture may occur along irregular, jagged, or splintery surfaces or along smooth, curved surfaces in what is called **conchoidal fracture** (Figure 3-9). Quartz is a common mineral that displays a striking conchoidal fracture. Ordinary glass also shows conchoidal fracture.

Luster

A mineral's **luster** is the manner in which light is reflected from its surface. Luster depends on the

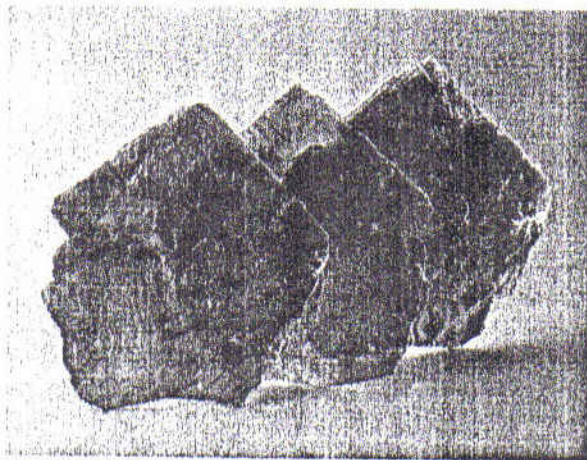


FIGURE 3-7

Perfectly developed cleavage in one direction shown by pieces of the mica called muscovite. Macmillan Publishing/Geoscience Resources photo.

composition of a mineral, the presence of defects or chemical impurities in its structure, and pitting or chemical alteration of its outer surfaces. Pure, unflawed diamonds have a *brilliant* luster, but inclusions of tiny impurities or flaws in the internal structure can cause a cloudy appearance. Most terms used to describe luster are self descriptive. Galena, the principal ore of lead, has a *metallic* luster—its surface looks like shiny aluminum foil. Other minerals have lusters that are silky, waxy, earthy, pearly, or glassy.

Specific Gravity

The **specific gravity** of a mineral is a comparison of

its density with the density of water, as shown in the relationship:

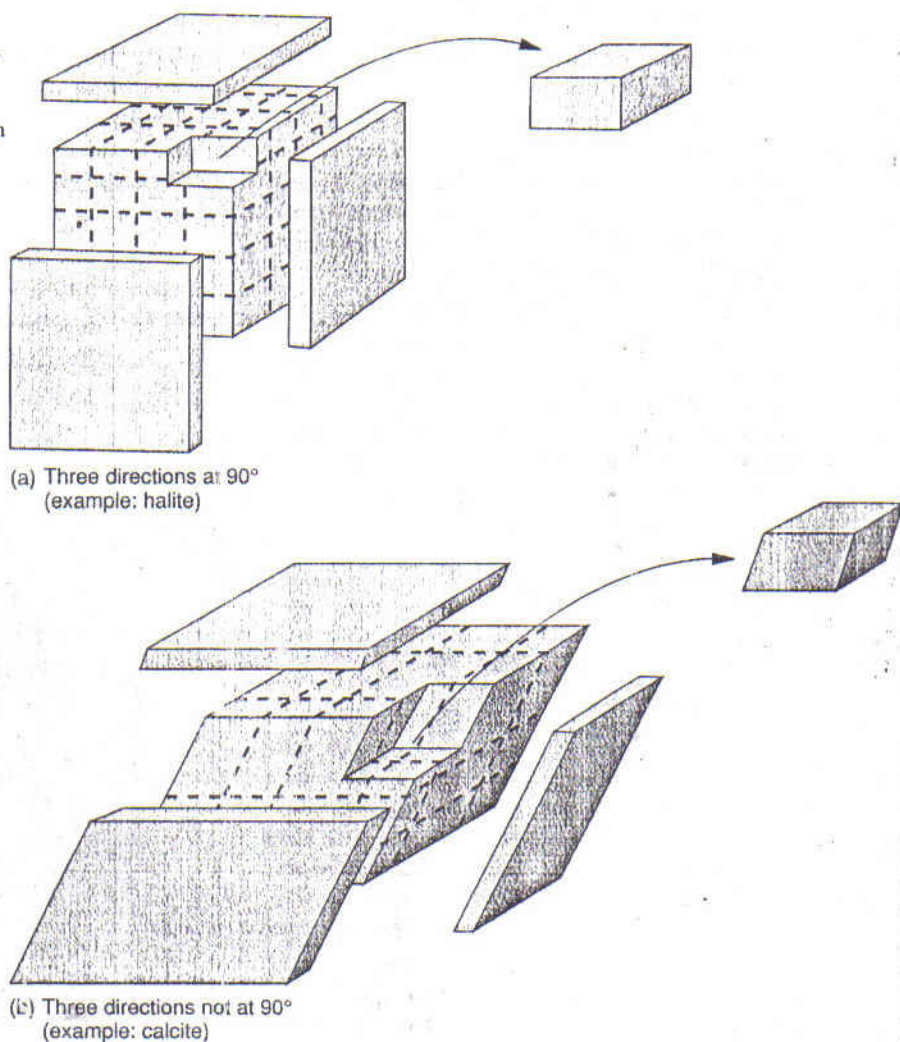
$$\text{Specific gravity} = \frac{\text{Density of a mineral}}{\text{Density of water}}$$

Density depends on the mass of the atoms in a mineral and the compactness with which they are bonded. Water at 25°C has a density of 1 g/cm³, and graphite has a density of 2.4 g/cm³. Graphite thus is 2.4 times as dense as water, so its specific gravity is 2.4—and if placed in water, graphite sinks.

If the specific gravity of a substance is drastically higher (or lower) than that of most materials, we

FIGURE 3-8

Angles between cleavage directions. Both halite and calcite have three directions of cleavage, but the angular relationships are different in the two minerals.



detect it when we hold the substance in our hands. If it seems much heavier (or lighter) than what we expect for the size of the sample, its specific gravity is abnormally high (or low). To determine the specific

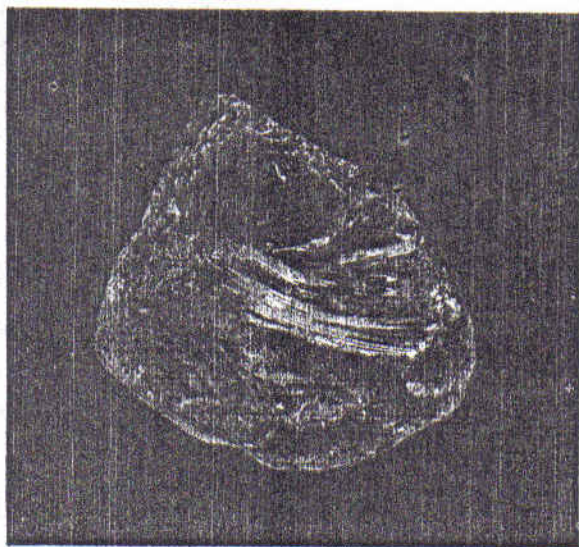
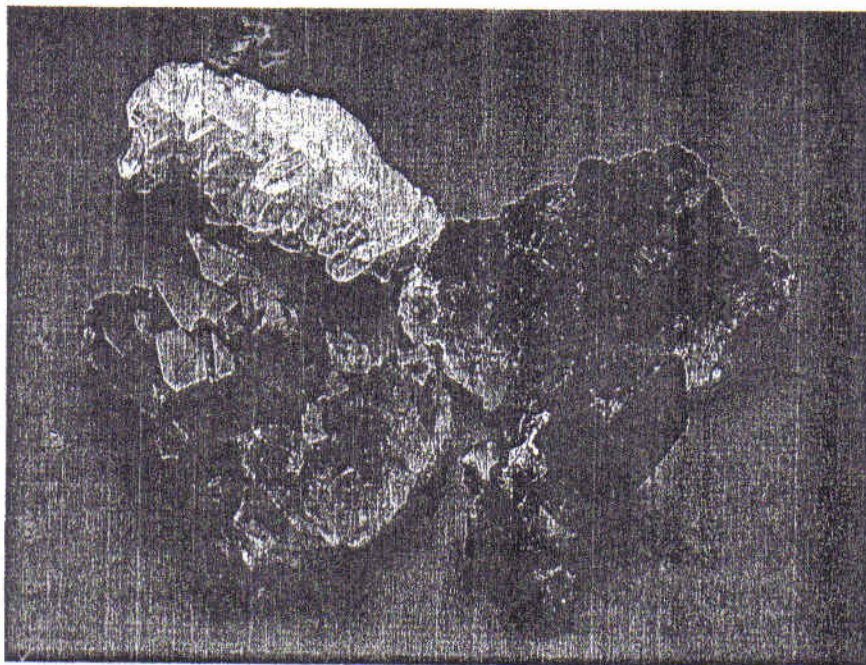


FIGURE 3-9
Conchoidal fracture exhibited by a quartz crystal. Macmillan Publishing/Geoscience Resources photo.

FIGURE 3-10
The color of a mineral may vary widely, as in these examples of quartz. Clockwise from upper left: colorless quartz, citrine (yellowish quartz), smoky quartz, and amethyst (purple quartz). Macmillan Publishing/Geoscience Resources photo.



gravity of a material whose density feels more in the normal range, we use a simple procedure: weigh a sample of the substance and compare its weight with that of an equal volume of water. The specific gravity is the weight of the material divided by the weight of the equal volume of water.

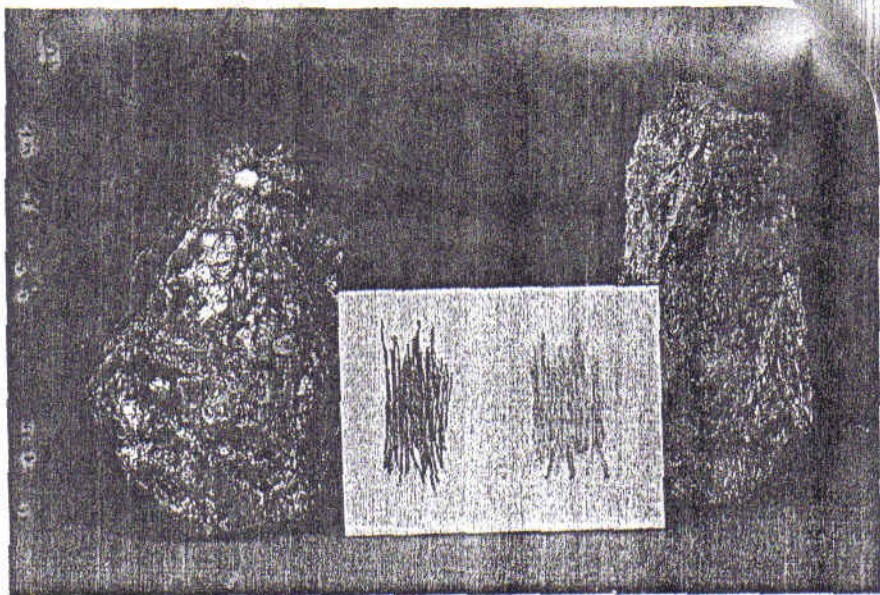
Most of the Earth's crust is composed of minerals that have specific gravities between 2.40 and 4.50, but some minerals have much higher values. Gold (15.0 to 19.3) and platinum (14.0 to 19.0) are the highest. Common ore minerals like galena (7.5) also are much denser than the usual crustal materials.

Color

Color is a useful property in identifying some minerals, but is much less valuable for others. Minerals with metallic lusters generally have colors that are diagnostic. Galena, for instance, is always battleship gray. Pyrite, also known as "fool's gold," is brassy yellow. Unfortunately, impurities and crystal defects cause many nonmetallic minerals to vary widely in color (Figure 3-10). Thus, quartz can be colorless, white, green, blue, gray, yellow, black, rose, smoky, or purple. Color *can be* a useful property, but it must be used with care.

FIGURE 3-11

Color versus streak. For some minerals, the streak is the same as the mineral color, but for other minerals the two colors are different. The yellow streak of *orpiment* (right) is the same as the color of the mineral. However, the brassy yellow *pyrite* (left) produces a black streak. Macmillan Publishing/Geoscience Resources photo.



Streak

The color of a mineral's powder is called its **streak**. It is so-called because it is determined by rubbing the mineral on a hard, white porcelain plate (hardness about 7), which grinds the mineral to a powdery streak on the plate. Streak is a more reliable property than surface color because pulverizing a mineral minimizes the effects of impurities, tarnish, or structural defects. For many minerals, color and streak are the same, but for many others they are surprisingly different, and this difference often can be a key to identification (Figure 3-11). (The few minerals that are harder than the streak plate leave no streak, of course—they scratch the plate.)

Unlike color, streak tends to stay constant. Thus, sphalerite, the major ore of zinc, has a creamy yellow streak regardless of whether the mineral is colorless, black, white, or green. If a brassy yellow mineral gives a black streak, chances are it is pyrite or another type of “fool’s gold.” However, if it streaks brassy or golden yellow, save the powder—it probably is real gold!

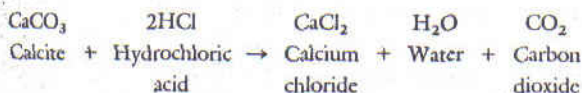
Other Properties

The preceding properties are easily determined and are sufficient for identification of many common

minerals, but several other properties also may prove helpful. Geologists often taste, sniff, and rub their fingers across mineral samples because taste, odor, and feel are diagnostic properties for certain minerals. Halite obviously tastes salty; kaolinite, a common mineral in clay, sticks to the tongue. The streak of many sulfur-bearing minerals smells like rotten eggs; that of some arsenic minerals smells like garlic. Graphite and talc feel greasy. Apply these tests carefully; geologists sniff first, to test for arsenic, before we taste!

Some minerals have one color in normal light, but look very different in ultraviolet light, a property called **fluorescence** (Figure 3-12). Other useful properties include **magnetism** (the ability of a mineral to attract a magnet), **malleability** (the ability of a mineral to be pounded into thin sheets), and **ductility** (the ability to be pulled into thin wires).

One simple chemical test often is made because it requires no equipment and only a drop or two of dilute hydrochloric acid. Minerals such as calcite that contain the carbonate anion, $(\text{CO}_3)^{2-}$, effervesce (“fizz”) when hydrochloric acid is put on them. A rapid chemical reaction takes place between the carbonate mineral and acid, releasing carbon dioxide and water. The carbon dioxide bubbles out through the liquid to produce the “fizz.” In the case of calcite, the reaction is:



MINERALS AND ROCKS

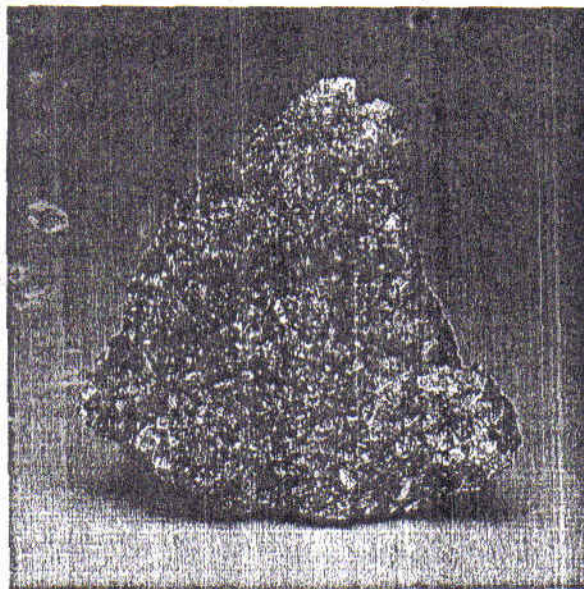
You may be wondering what the difference is between a mineral and a rock. A **rock** is an aggregate of several different minerals or of many grains of a single mineral. It is rare for a mineral to grow in the isolation needed to form a perfect crystal. Most minerals are intergrown with many other mineral grains in a wide variety of rocks.

How Do Minerals Form?

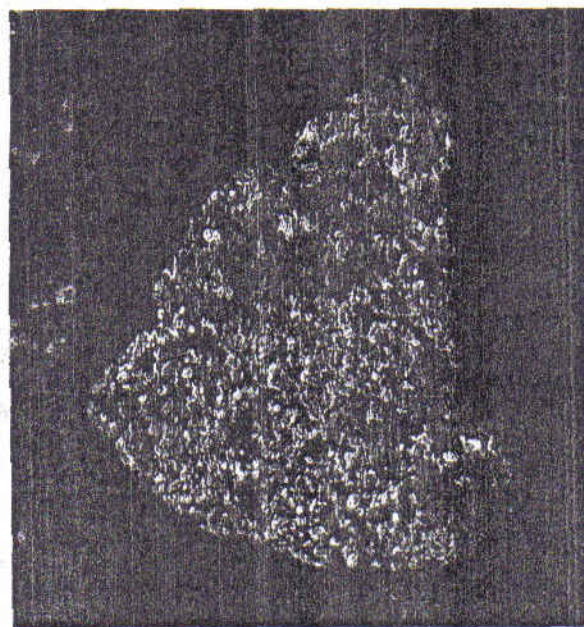
Minerals grow from small clusters of atoms or ions. At first, a small number of ions forms a tiny **seed crystal**—a group of ions in the appropriate proportions of elements and arranged in the correct internal structure. More ions are attracted to oppositely charged ions in the seed crystal. As the additional ions fit into their structural positions, the seed grows into a visible grain. If seed crystals are far apart, each can grow without interference from the others, and well-formed crystals are produced (see Focus 3–1). If several seed crystals are close to one another, they interfere with each other as they grow, resulting in an interlocking aggregate of grains that displays no crystal faces.

Minerals grow in a wide range of physical conditions, some from liquids and others from solids. Ice forms by the solidification of water at 0°C . At slightly higher temperatures, halite forms by precipitating from saltwater as the water evaporates. Other minerals, such as olivine and feldspar, grow deep within the crust from molten rock at temperatures of 550° to 1100°C and pressures up to 10 kb. Still others, such as garnet, form in solid rocks at moderate temperatures (350° – 600°C) and pressures (2–8 kb) by the rearrangement of ions already present in other minerals. (You can even grow mineral-like crystals yourself—see Focus 3–1.)

Some minerals, such as quartz, can grow under a wide range of conditions, from those of the surface to those several kilometers deep in the crust. Others, such as halite, form only within a restricted range of conditions. Those that form in a narrow range of



(a)



(b)

FIGURE 3–12

Some minerals *fluoresce*; that is, they exhibit different colors when viewed under ultraviolet light. Calcite (white), willemite, and franklinite in this specimen from Franklin, New Jersey, are pale in normal light (a) but fluoresce brightly in ultraviolet light (b). (The willemite fluoresces green.) Macmillan Publishing/Geoscience Resources photos.

conditions are used by geologists to estimate the conditions under which the rocks that contain them formed.

Rocks and Rock-Forming Processes

Rocks are divided into three types—*igneous*, *sedimentary*, and *metamorphic*—according to the nature of the processes that brought the minerals together. Minerals also can grow during each of the three processes and thus are igneous, sedimentary, or metamorphic minerals. We will look briefly at these processes here, and study them in more detail in Chapters 4, 5, 6, and 7.

Molten rock is called **magma**, and rocks that form by the crystallization of magma are known as **igneous rocks**. Seed crystals form in a magma as it cools, and the igneous minerals grow in the hot liquid. Eventually, all of the ions in the liquid fit into crystal structures and the magma is completely solidified as an igneous rock. Igneous minerals and rocks form deep in the crust and mantle and at the surface during volcanic eruptions. Olivine, feldspars, and quartz can crystallize from magmas and thus may be igneous minerals.

When rocks are exposed at the surface of the Earth, they eventually crumble or dissolve. This breakdown is caused partly by chemical reactions with the atmosphere and partly by physical processes such as wedging caused by ice and plant roots. Fragmented and dissolved rock material can be transported by streams, glaciers, and wind, then deposited or precipitated from solution, and finally cemented together to produce a new rock, such as sandstone. Rocks produced at the surface by these processes are called **sedimentary rocks**.

Sedimentary minerals are the new minerals that grow chemically during the deposition or cementation processes, usually by precipitation of dissolved ions from water. Halite, for example, is a sedimentary mineral formed by evaporation of seawater or saline lakes in arid areas. Some quartz also is sedimentary because it precipitates between fragments and acts as a cement.

Note that some sedimentary rocks may be composed of reworked fragments of any kind of previously existing rock, not just of new sedimentary minerals. Thus, old minerals—igneous, sedimentary, or

metamorphic—may be found in sedimentary rocks.

When igneous and sedimentary rocks are subjected to temperatures and pressures greatly different from those under which they first formed, they often respond by forming new minerals. These minerals form without melting or dissolving, through the recombination of ions from previously existing minerals. The new minerals are **metamorphic minerals**, and the resulting rocks are called **metamorphic rocks**. Common examples of metamorphic minerals are those of the garnet group.

Many minerals, such as halite, can form only by one of the three types of rock-forming processes, but others, such as olivine, may form by two processes (igneous and metamorphic). A few minerals can form during all three kinds of processes. For example, quartz and some of the feldspars may be igneous, sedimentary, or metamorphic minerals.

The Rock Cycle

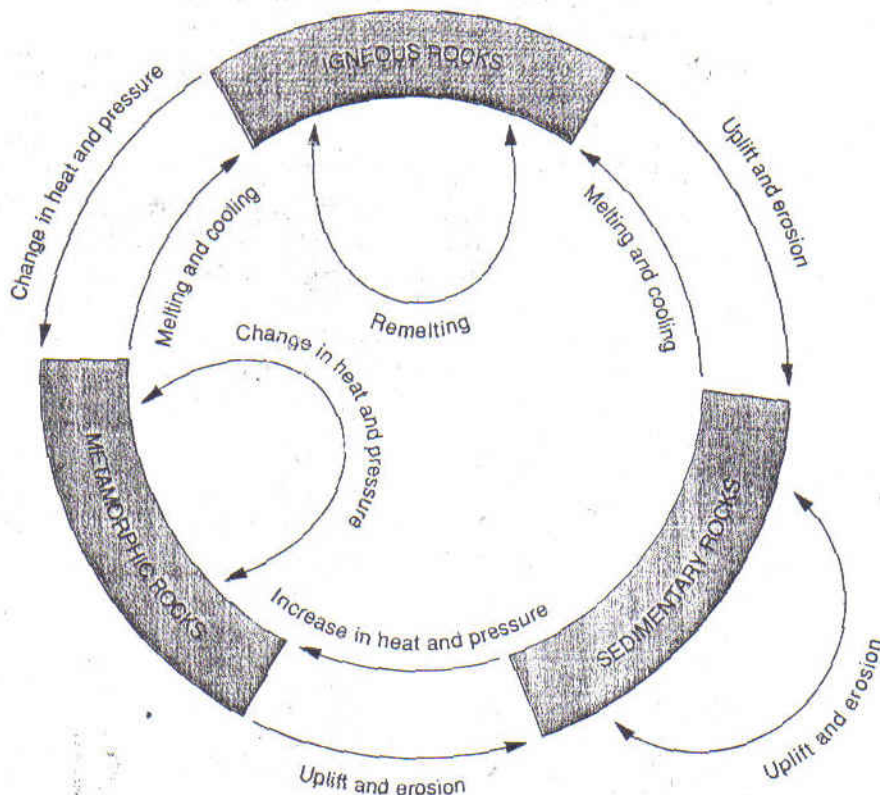
Any rock can be converted to an igneous, sedimentary, or metamorphic rock if it is acted on by the appropriate processes. Nature is thus the original recycler, using the same ions over and over again to make new rocks in another important Earth cycle called the **rock cycle** (Figure 3-13). The rock cycle is more complex than the water cycle because it uses deep internal processes as well as those that operate at the surface.

One way to show how the rock cycle works is to look at what happens when lithospheric plates collide. When sedimentary rocks formed on the ocean floor are subducted, heat and pressure first convert them to metamorphic rocks as they are buried deep in the crust. Eventually, they reach a level in the mantle where they melt to form magma. This magma rises and becomes igneous rock by solidifying deep in the crust or after erupting at the surface. Breakdown of lava at the surface converts the igneous rock back to sedimentary rocks, completing the cycle.

The rock cycle follows other paths as well. Sedimentary, metamorphic, and igneous rocks exposed at the surface will be converted to new sedimentary rocks, or they will be converted to metamorphic rocks if they are buried beneath new deposits. Metamorphic rocks may be brought to the surface during mountain building and converted to sediments, or

FIGURE 3-13

The rock cycle. Depending on the processes to which it is subjected, any rock may be converted to an igneous, sedimentary, or metamorphic rock. Arrows show the possible conversions.



they may be melted during subduction. With deeper burial, they also may be remetamorphosed into a new metamorphic rock. An igneous rock also can remelt and become a different igneous rock! Figure 3-13 summarizes the many possible paths that the rock cycle can take. We will be looking at the processes that occur along those paths throughout the rest of this book.

MINERALS OF THE EARTH'S CRUST

There are nearly 3000 known minerals, and new ones are found every year. Fortunately for students, only a few of the 3000 are abundant; we need to know only about these few to understand the processes by which igneous, sedimentary, and metamorphic rocks form. These few are called the **rock-forming minerals**. We will examine some of the most important groups in the rest of this chapter.

Minerals are classified by the anion they contain. For example, those that contain the sulfur anion are

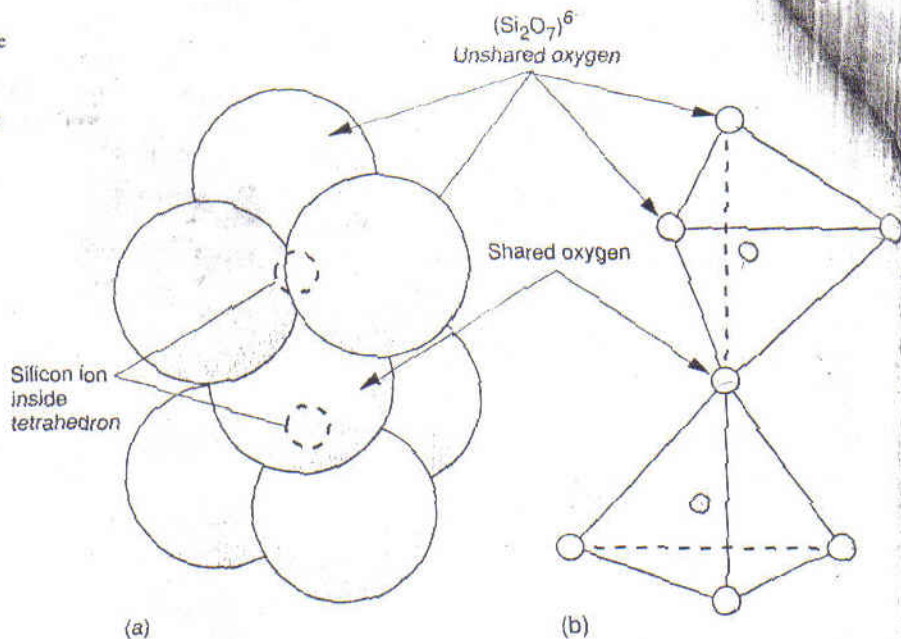
sulfides, and those containing chlorine are called chlorides. We saw in Chapter 2 that the crust is composed predominantly of oxygen and silicon. Thus, it is not surprising that the Earth's crust consists mostly of minerals containing *both* elements—minerals called **silicates**. The rock-forming minerals are generally divided into two major groups: silicates and nonsilicates.

Silicate Minerals

The relative sizes of the Si^{4+} and O^{2-} ions are such that four oxygens can fit around a silicon ion. They form a four-sided figure called a **silicon-oxygen tetrahedron**, with silicon at the center and the oxygens at the corners—Figure 3-5(b). The SiO_4 tetrahedron is not electrically neutral; the charge of the four oxygens adds to -8 , and that of the silicon is $+4$. Consequently, the SiO_4 tetrahedron acts as an **anion complex** with a total charge of -4 . Importantly, an SiO_4 tetrahedron *cannot exist by itself* because it is not neutral and must attract cations to achieve neu-

FIGURE 3-14

The twin-tetrahedron structure. The $(\text{Si}_2\text{O}_7)^{6-}$ anion complex forms when two tetrahedra share an oxygen ion. (a) Proportion of ionic sizes. (b) Stick model of twin tetrahedron showing oxygens at corners and silicons at centers.



trality. These cations hold the tetrahedra together and stabilize the silicate mineral structures.

The silicon-oxygen tetrahedron is the basic building block of all silicate minerals, but building blocks can be put together in many ways. A brief look at the silicate minerals will give more insight into the variety of mineral structures.

Types of Silicate Minerals. The simplest silicate structure is one in which SiO_4 tetrahedra are held together by cations. This type of structure is called an **independent tetrahedron structure** and is found in several minerals, including those of the olivine family— $(\text{Fe}, \text{Mg})_2\text{SiO}_4$. Each oxygen gives some of its negative charge to a silicon ion and some to iron ions or magnesium ions. The tetrahedra must be arranged in exactly the right way so that their oxygens also fit into the correct coordination around the iron and magnesium.

For independent tetrahedra to form, there must be four oxygens for every silicon. Indeed, every mineral that has the independent tetrahedron structure has SiO_4 in its formula, reflecting this ratio. If there are not enough oxygens to go around during rock-forming processes, independent tetrahedra cannot exist and a different kind of structure must be

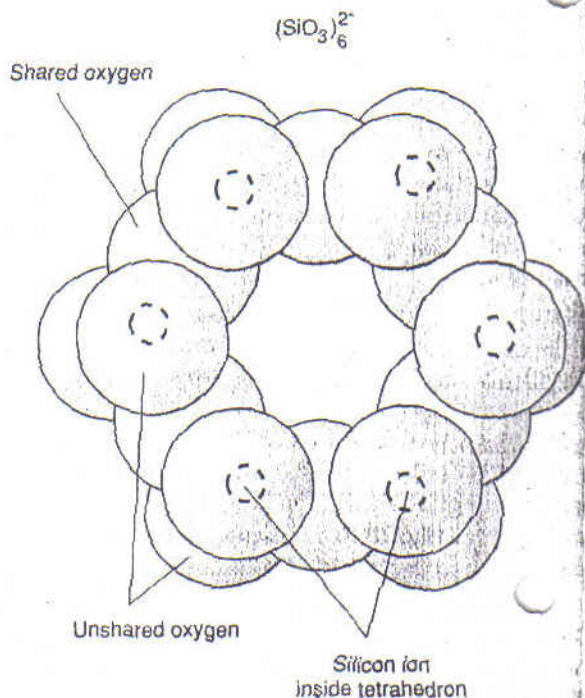


FIGURE 3-15

Geometry of the ring silicate structure. Each tetrahedron shares two oxygens with neighboring tetrahedra. In this case, the $(\text{Si}_6\text{O}_{18})^{12-}$ anion complex is formed.

formed. In such a case, the available oxygens must be *shared* between tetrahedra.

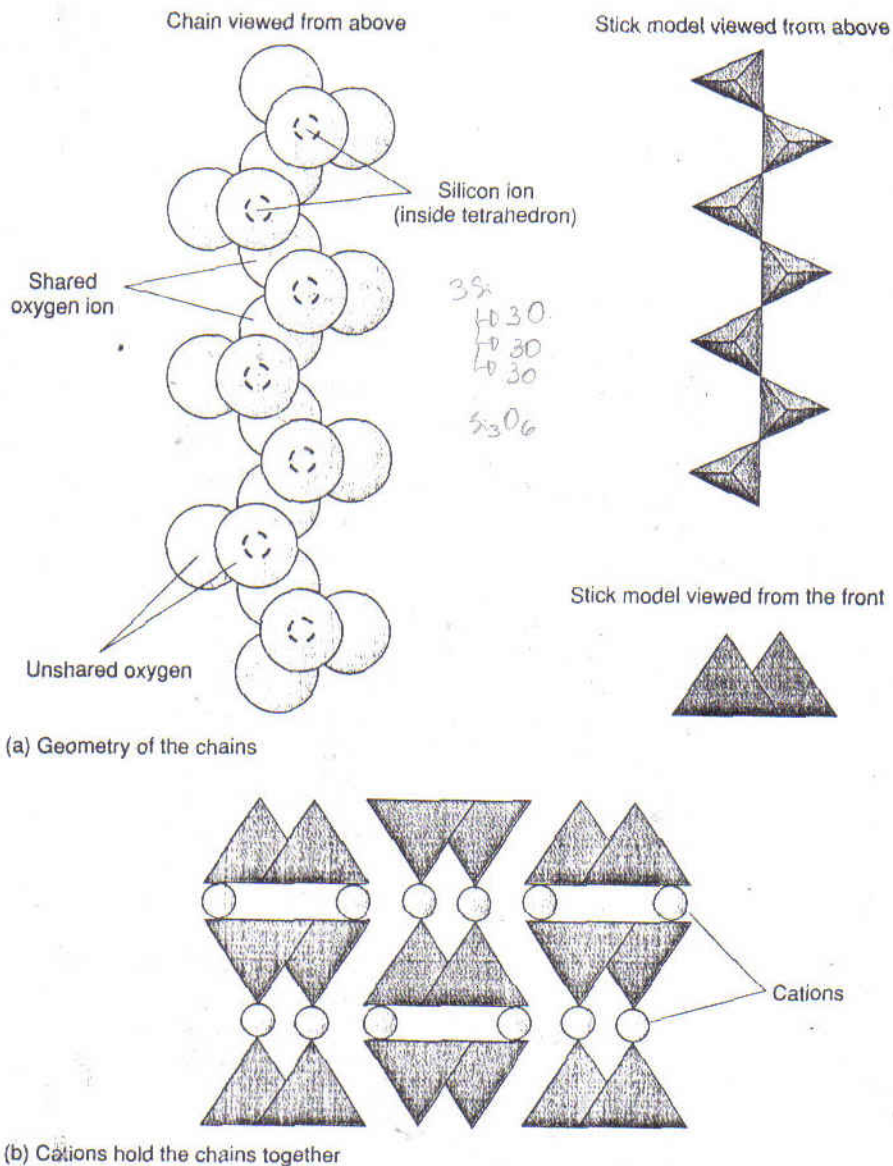
In some minerals, two tetrahedra share an oxygen ion and are linked in a **twin-tetrahedron structure** (Figure 3-14). Notice that each Si^{4+} ion is surrounded by four oxygens, as required by the coordination principle, but that one of the oxygens (the *shared oxygen*) is part of *both* tetrahedra. The twin tetrahedron has a composition of Si_2O_7 and a net charge of -6 ; it acts as an anion complex and attracts cations to satisfy its remaining charge.

In other silicate minerals, each tetrahedron shares two of its four oxygens with adjacent tetrahedra. This can be accomplished in two ways. In the **ring silicates**, rings made of three, four, or six tetrahedra are formed, and *these rings* become the anion complexes that act as building blocks in the mineral structure (Figure 3-15). The six-tetrahedron ring shown in Figure 3-15 is typical of the minerals tourmaline and beryl. In other minerals, the tetrahedra line up to form long chains by sharing oxygens with their neighbors in what is called a **single-chain structure**

FIGURE 3-16

The single-chain silicate structure.

(a) Geometry of the chain shows each tetrahedron sharing two oxygens with its neighbors along the chain. The shared oxygens are electrically satisfied by being bonded to two silicon ions. (b) Unshared oxygens retain a negative charge and attract the cations that hold the chains together.

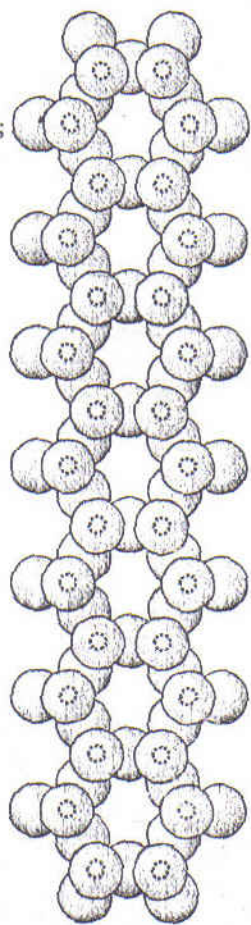


(Figure 3-16). These chains act as large anion complexes and are held together by cations. In both the ring and single-chain structures, the Si:O ratio is 1:3.

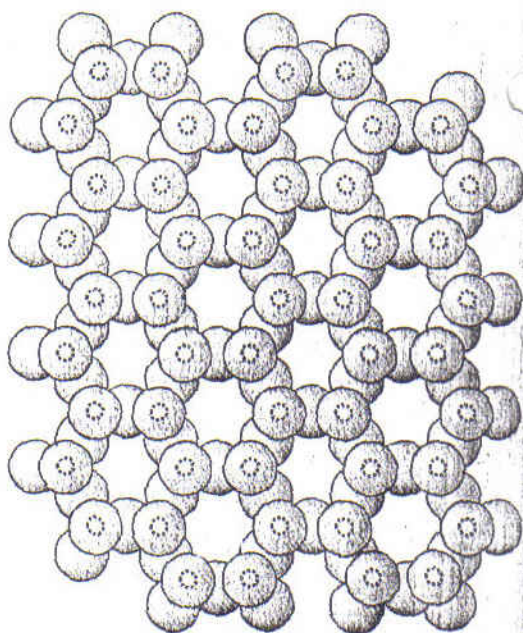
With progressively smaller amounts of oxygen, even more sharing must take place, resulting in even more-elaborate anion complexes. **Double chains** form when some tetrahedra share two oxygens and others share three, resulting in an Si:O ratio of 4:11—Figure 3-17(a). **Sheet silicate structures** with a ratio of 2:5 occur when every tetrahedron shares three of its four oxygens with its neighbors—Figure 3-17(b). This sheet structure is reflected clearly in the sheetlike cleavage found in the mica minerals. When all four oxygens of every tetrahedron are shared with neighboring tetrahedra, a complex three-dimensional network called a **framework silicate structure** is formed (Figure 3-18).

FIGURE 3-17

Geometry of (a) double-chain and (b) sheet silicate structures. In double chains, some tetrahedra share three oxygens with neighbors; others share four. In sheets, all tetrahedra share three oxygens.



(a) Double chain (Si_4O_{11})



(b) Sheet (Si_2O_5)

Three of the most abundant minerals in the Earth's crust have framework structures—quartz, plagioclase feldspar, and potassic feldspar.

Rock-Forming Silicate Minerals. The olivine group consists of the solid-solution series forsterite-fayalite and has the independent tetrahedron structure. Olivine minerals are typically green and have high specific gravities for silicate minerals (3.27 to 4.37). Most olivine occurs in stubby, irregularly shaped grains. As a result, rocks composed entirely of olivine are compact masses lacking the elongate crystals common in rocks containing feldspars or amphiboles. Olivines form throughout the crust in igneous rocks, but the magnesium-rich varieties are found also in metamorphic rocks. Olivine is thought to be a major constituent of the mantle and has been found in

many meteorites. Specimens with particularly brilliant luster are valued as the gemstone peridot.

The **garnet family** consists of two solid-solution series: $(\text{Fe}, \text{Mg}, \text{Mn})_3\text{Al}_2(\text{SiO}_4)_3$ and $\text{Ca}_3(\text{Al}, \text{Fe}, \text{Cr})_2(\text{SiO}_4)_3$. Each has three end members. With all of the substitutions that are possible, it is not surprising that garnets come in several colors. Dark red is the most common, but pink, brown, yellow, and bright green types also are known. Garnets are independent tetrahedron silicates; most form in metamorphic rocks, although a small proportion crystallize in igneous rocks.

The **pyroxene family** consists of several solid-solution series, all of which have the single-chain structure. The weakest bonds in the pyroxene structure lie between the chains and result in the characteristic two-directional cleavage at nearly right angles (87° and 93°). Pyroxenes are commonly medium-to-dark green or bronze-brown. One type, known as jadeite, has a color and unique luster that makes it valuable as the familiar gemstone "jade."

Pyroxenes are important constituents of igneous and metamorphic rocks and are thought to be important components of the mantle. The most com-

mon pyroxenes belong to one of two solid-solution series: the complex augite group, $(\text{Ca}, \text{Na})(\text{Mg}, \text{Fe}, \text{Al})(\text{Si}, \text{Al})_2\text{O}_6$, and the simpler enstatite group, $(\text{Mg}, \text{Fe})\text{SiO}_3$.

The **amphibole family** consists of several solid-solution series similar to those of the pyroxenes, but with a double-chain structure in which the hydroxyl anion complex (OH^-) is present. Amphiboles are difficult to distinguish from pyroxenes because both are chain silicates, contain the same elements, are the same color, and occur in the same kinds of rocks. In many instances, the distinction can be made only by the different cleavages of these two mineral families: amphiboles also cleave between their chains, but at angles of 57° and 123° rather than at right angles.

Amphiboles are important igneous and metamorphic minerals, the most common belong to the hornblende "series," actually a collection of over ten solid-solution series. Hornblendes form under a wide range of conditions, but other amphiboles are more restricted in their occurrence.

The **mica family** consists of several sheet silicate minerals in which large cations—usually K^+ , Na^+ , or Ca^{2+} —bond silicon-oxygen sheets together. The most-abundant micas are the colorless mineral muscovite and the dark brown or green mineral biotite. Both are important constituents of metamorphic rocks and form in many kinds of igneous rock as well. All micas have such excellent cleavage in one direction (between the silicon-oxygen sheets) that they can be peeled into thin, flexible sheets (Figure 3-7).

The **feldspars** include two groups of minerals, the *plagioclase* and *potassic feldspar* families. Each has a framework structure in which aluminum ions substitute for silicon in some tetrahedra, and large cations are added to the structure in a coupled ionic substitution.

The **plagioclase feldspars** are a solid-solution series of framework silicates varying in composition between end members albite, $\text{NaAlSi}_3\text{O}_8$, and anorthite, $\text{CaAl}_2\text{Si}_2\text{O}_8$. Plagioclase minerals are the most abundant in the Earth's crust and are found in nearly every igneous and metamorphic rock. In addition, sodium-rich plagioclase (albite) can form in some sedimentary environments.

Plagioclases vary widely in color, occurring in white, gray, black, and colorless varieties, but can be recognized by their good two-directional cleavage at

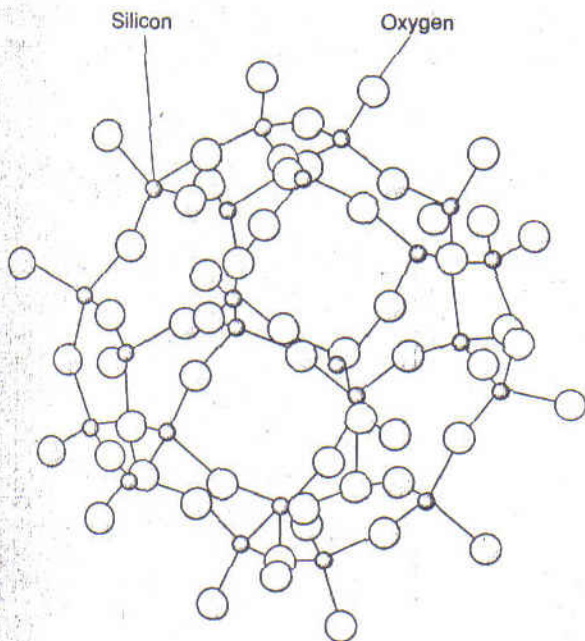


FIGURE 3-18

Part of a framework silicate structure. All four oxygen ions of each tetrahedron are shared with neighboring tetrahedra in this complex three-dimensional network.

nearly right angles and their moderate hardness (6 on Mohs scale). Some plagioclase grains exhibit very fine grooves called *striations* on one of the two cleavage directions; these striations are a diagnostic property for the entire family.

The **potassic feldspars** also are framework silicates and have the composition KAlSi_3O_8 . There are three polymorphs of KAlSi_3O_8 , each of which forms in different environments and can help pinpoint the processes or conditions of formation. **Sanidine** crystallizes in volcanic igneous rocks and in some metamorphic rocks that have experienced very high temperatures and low pressures. **Orthoclase** forms under higher-pressure metamorphic conditions, in igneous rocks, and to a lesser extent in sedimentary rocks. The third polymorph, **microcline**, forms in igneous and metamorphic environments similar to those of orthoclase, but is not produced in sedimentary processes.

Quartz, SiO_2 , is one of the commonest minerals and can form in igneous, metamorphic, and sedimentary rocks. It is a framework silicate, but unlike the feldspars there is no substitution of aluminum for silicon. Quartz is identified readily in rocks by its glassy luster, high hardness (7), and conchoidal fracture. Impurities and defects in its crystalline structure help quartz to occur in many colors. Its colorless

variety is known as rock crystal, and each of its colored varieties has been given a different name: amethyst (purple), rose quartz (pink), smoky quartz (gray or black), citrine (yellow), and aventurine (green).

Nonsilicate Minerals. Although there are more *kinds* of nonsilicate minerals than silicates, the nonsilicates constitute only a very small part of the Earth's crust. Few nonsilicates are important rock-forming minerals; most occur as **accessory minerals** (minor minerals) in silicate rocks. On the other hand, nonsilicates are very important economically because they include most of the ore minerals from which we get metals and nonmetallic resources. A list of some of the more important nonsilicate minerals is given in Table 3–3. Our discussion here will focus on those few nonsilicates that are rock-forming minerals (highlighted in Table 3–3).

There are several **carbonate minerals**, each of which contains the carbonate (CO_3)²⁻ anion complex. The two most common are **calcite**, CaCO_3 , and **dolomite**, $\text{CaMg}(\text{CO}_3)_2$. These are important constituents of sedimentary rocks, but also are abundant in some metamorphic rocks. Both are relatively soft (around 3 on Mohs scale) and display excellent cleavage in three nonperpendicular directions. Both

TABLE 3–3
Important nonsilicate minerals. Rock-forming minerals are boldfaced.

Native Elements	Oxides (O^{2-})	Sulfides (S)
Gold (Au)	Hematite: Fe_2O_3	Galena: PbS
Silver (Ag)	Magnetite: $\text{FeO} \cdot \text{Fe}_2\text{O}_3$	Pyrite: FeS_2
Platinum (Pt)	Corundum: Al_2O_3	Bornite: Cu_5FeS_4
Diamond (C)	Ilmenite: FeTiO_3	Sphalerite: ZnS
Sulfur (S)	Chromite: FeCr_2O_4	
Carbonates (CO_3)	Halides (Cl^- , F^-)	Sulfates (SO_4) ²⁻
Calcite: CaCO_3	Halite: NaCl	Gypsum: $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$
Aragonite: CaCO_3	Sylvite: KCl	Anhydrite: CaSO_4
Dolomite: $\text{CaMg}(\text{CO}_3)_2$	Fluorite: CaF_2	Barite: BaSO_4
Malachite: $\text{Cu}_2(\text{CO}_3)(\text{OH})_2$		
Azurite: $\text{Cu}_3(\text{CO}_3)_2(\text{OH})_2$		
Hydroxides (OH^-)	Phosphates (PO_4) ³⁻	Others
Goethite: $\text{FeO}(\text{OH})$	Apatite: $\text{Ca}_5(\text{F,Cl,OH})(\text{PO}_4)_3$	Borax: $\text{Na}_2\text{B}_4\text{O}_7 \cdot 10\text{H}_2\text{O}$
Gibbsite: $\text{Al}_2(\text{OH})_6$	Monazite: $(\text{Ce,La,Y,Th})\text{PO}_4$	Carnotite: $\text{K}_2(\text{UO}_2)_2(\text{VO}_4)_2 \cdot 3\text{H}_2\text{O}$
Boehmite: $\text{AlO}(\text{OH})$	Turquoise: $\text{CuAl}_6(\text{PO}_4)_4(\text{OH})_6$	Wulfenite: PbMoO_4

react with hydrochloric acid, but differently: calcite effervesces vigorously and dolomite only weakly.

Gypsum, $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$, and anhydrite, CaSO_4 , are the most common minerals containing the sulfate anion complex (SO_4^{2-}). They are typically light gray to white and are among the softest minerals known (gypsum is the standard mineral for 2 on the Mohs scale). Both minerals occur in massive aggregates or well-developed crystals, some up to 1.5 m long. Gypsum and anhydrite are exclusively sedimentary in origin, and form by evaporation of seawater.

Elements next to the inert gases in the periodic table are called halogen elements, and their minerals are called **halides**. The only major halide rock-forming mineral is halite, NaCl , although fluorite, CaF_2 , is a very common accessory mineral. Halite occurs both in clear cubic crystals that exhibit cleavage in three mutually perpendicular directions and in granular masses that form layers tens of meters thick ("rock salt") in many areas. Like gypsum and anhydrite, halite is a sedimentary mineral formed by evaporation of seawater.

One important group of minerals—the clay minerals—does not fit into our simple chemical classification scheme because some clays are silicates and others are nonsilicates. All form at the Earth's surface by **weathering**, which is the interaction between previously formed minerals and the gases of the atmosphere. These minerals contain aluminum and various amounts of water and/or the hydroxyl anion complex in their structures. Clays commonly occur in very fine-grained mixtures that are plastic and can be molded when mixed with a little water.

The most common clay minerals are the silicates kaolinite, montmorillonite, and illite (a mineral much like muscovite but with less potassium and more silicon). Their silicon:oxygen ratios indicate that these clay minerals have the sheet silicate structure. Gibbsite and boehmite are the most common nonsilicate clays. All the clays contain aluminum and various amounts of water or the hydroxyl anion complex.

USES OF MINERALS

Many minerals are useful, and we go to extraordinary lengths to discover and mine them. Today we tunnel thousands of meters below the surface for copper

sulfides and diamonds, pump steam into the ground to melt and recover native sulfur, and even talk about dredging the deep ocean floors for manganese oxides. Minerals are valuable resources only if they are concentrated in such large amounts that recovery is easy. In Chapters 4, 6, and 7 we will look at some of the rock-forming processes that bring about these concentrations, and in Chapter 21 will look at all of our natural resources in some detail. Here, we will focus on those attributes of minerals that make them useful to humans. A list of a few of the important mineral resources, both metals and nonmetals, is given in Table 3-4.

Some minerals are valuable because of the *elements they contain*. The three metals we use most—iron, aluminum, and copper—come from mineral oxides (magnetite and hematite for iron), mineral hydroxides (bauxite for aluminum), and mineral sulfides (bornite for copper). Minerals containing uranium, such as carnotite and uraninite, are important sources of nuclear fuels.

Other minerals are useful because of their *physical properties*. Industrial abrasives used for smoothing, polishing, and grinding are chosen from minerals having great hardness—quartz (hardness 7), corundum (9), and diamond (10). The drills used to penetrate kilometers of rock in the search for oil and natural gas are coated with diamonds to help grind through the softer rock-forming minerals of the crust. Graphite, on the other hand, is so soft that it rubs off on paper; it is graphite that you write with when you use a common black pencil. (Actually, graphite is too soft to be used alone in pencils because it smudges so easily. It must be mixed with clay minerals to reach the desired hardness.)

Talc and graphite often are used as lubricants in industry because of their softness and greasiness. The greasy feel of these minerals is due to the presence of very weak van der Waals' bonds in their structures. Finger pressure alone can break these bonds, causing sheets of silicon-oxygen tetrahedra (in talc) or of carbon atoms (in graphite) to slide past one another. It is this sliding that we perceive as greasiness. In contrast, extremely strong bonds give very high melting points to feldspars and to the Al_2SiO_5 polymorphs named andalusite, sillimanite, and kyanite. These minerals are used in making porcelain and the refractory bricks that line the insides of steelmaking blast furnaces.