general agreement that the continental crust formed gradually over the last 4 billion years. (The oldest rocks yet discovered are isolated fragments found in the Northwest Territories of Canada that have radiometric dates of about 4 billion years.) In addition, as you will see in subsequent chapters, Earth is an evolving planet whose continents (and ocean basins) have continually changed shape and even location during much of this period.

CONCEPT CHECK 1.9

1. Name and briefly outline the theory that describes the formation of our solar system.
2. Name the inner planets and the outer planets. Describe basic differences in size and composition.

Earth's Internal Structure

An Introduction to Geology:
Earth's Layered Structure

In the preceding section, you learned that the segregation of material that began early in Earth's history resulted in the formation of three layers defined by their chemical composition—the crust, mantle, and core. In addition to these compositionally distinct layers, Earth can be divided into layers based on physical properties. The physical properties used to define such zones include whether the layer is solid or liquid and how weak or strong it is. Knowledge of both types of layered structures is essential to our understanding of basic geologic processes, such as volcanism, earthquakes, and mountain building (FIGURE 1.26).

Earth's Crust

The **crust**, Earth's relatively thin, rocky outer skin, is of two different types—continental crust and oceanic crust. Both share the word "crust," but the similarity ends there. The oceanic crust is roughly 7 kilometers (5 miles) thick and composed of the dark igneous rock *basalt*. By contrast, the continental crust averages about 35 kilometers (22 miles) thick

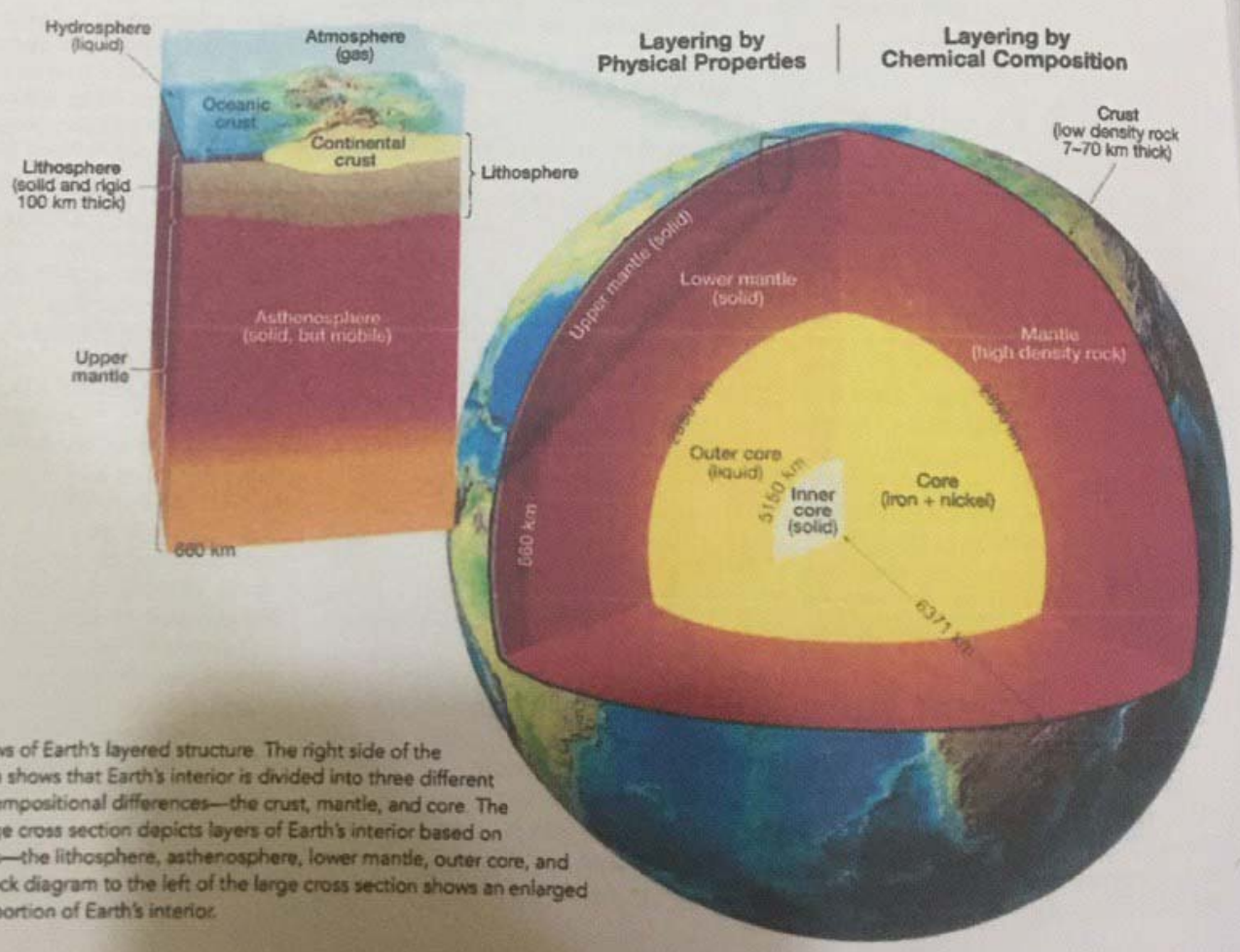


FIGURE 1.26 Views of Earth's layered structure. The right side of the large cross section shows that Earth's interior is divided into three different layers based on compositional differences—the crust, mantle, and core. The left side of the large cross section depicts layers of Earth's interior based on physical properties—the lithosphere, asthenosphere, lower mantle, outer core, and inner core. The block diagram to the left of the large cross section shows an enlarged view of the upper portion of Earth's interior.

but may exceed 70 kilometers (40 miles) in some mountainous regions such as the Rockies and Himalayas. Unlike the oceanic crust, which has a relatively homogeneous chemical composition, the continental crust consists of many rock types. Although the upper crust has an average composition of a *granitic rock* called *granodiorite*, it varies considerably from place to place.

Continental rocks have an average density of about 2.7 g/cm^3, and some have been discovered that are 4 billion years old. The rocks of the oceanic crust are younger (180 million years or less) and denser (about 3.0 g/cm^3) than continental rocks.*

Earth's Mantle

More than 82 percent of Earth's volume is contained in the **mantle**, a solid, rocky shell that extends to a depth of nearly 2900 kilometers (1800 miles). The boundary between the crust and mantle represents a significant change in chemical composition. The dominant rock type in the uppermost mantle is *peridotite*, which is richer in the metals magnesium and iron than the minerals found in either the continental or oceanic crust.

THE UPPER MANTLE. The upper mantle extends from the crust-mantle boundary down to a depth of about 660 kilometers (410 miles). The upper mantle can be divided into two different parts. The top portion of the upper mantle is part of the stiff *lithosphere*, and beneath that is the weaker **asthenosphere.**

The **lithosphere** (sphere of rock) consists of the entire crust and uppermost mantle and forms Earth's relatively cool, rigid outer shell. Averaging about 100 kilometers in thickness, the lithosphere is more than 250 kilometers thick below the oldest portions of the continents. Beneath this stiff layer to a depth of about 350 kilometers lies a soft, comparatively weak layer known as the *asthenosphere* ("weak sphere"). The top portion of the asthenosphere has a temperature/pressure regime that results in a small amount of melting. Within this weak zone the lithosphere is mechanically detached from the layer below. The result is that the lithosphere is able to move independently of the asthenosphere, a fact we will consider later in the chapter and in Chapter 15.

It is important to emphasize that the strength of various Earth materials is a function of both their composition and the temperature and pressure of their environment. You should not get the idea that the entire lithosphere behaves like a brittle solid similar to rocks found on the surface. Rather, the rocks of the lithosphere get progressively hotter and weaker (more easily deformed) with increasing depth. At the depth of the uppermost asthenosphere, the rocks are close enough to their melting temperature (some melting may actually occur) that they are very easily deformed. Thus, the uppermost asthenosphere is weak because it is near its melting point, just as hot wax is weaker than cold wax.

THE LOWER MANTLE. From a depth of 660 kilometers (nearly 410 miles) to the top of the core, at a depth of 2900 kilometers (1800 miles), is the **lower mantle.** Because of an increase in pressure (caused by the weight of the rock above) the mantle gradually strengthens with depth. Despite their strength however, the rocks within the lower mantle are very hot and capable of very gradual flow.

Earth's Core

The composition of the **core** is thought to be an iron-nickel alloy with minor amounts of oxygen, silicon, and sulfur—elements that readily form compounds with iron. At the extreme pressure found in the core, this iron-rich material has an average density of nearly 11 g/cm^3 and approaches 14 times the density of water at Earth's center.

The core is divided into two regions that exhibit very different mechanical strengths. The **outer core** is a *liquid layer* 2270 kilometers (1410 miles) thick. It is the movement of metallic iron within this zone that generates Earth's magnetic field. The **inner core** is a sphere having a radius of 1216 kilometers (754 miles). Despite its higher temperature, the iron in the inner core is *solid* due to the immense pressures that exist in the center of the planet.

DID YOU KNOW?

We have never sampled the mantle or core directly. The structure of Earth's interior is determined by analyzing seismic waves from earthquakes. As these waves of energy penetrate Earth's interior, they change speed and are bent and reflected as they move through zones having different properties. Monitoring stations around the world detect and record this energy.

CONCEPT CHECK 1.10

1. List and briefly describe Earth's compositional layers.
2. Contrast the lithosphere and the asthenosphere.

The Face of Earth

The two principal divisions of Earth's surface are the continents and the ocean basins (**FIGURE 1.27**). A significant difference between these two areas is their relative levels. The continents are remarkably flat features that have the appearance of plateaus protruding above sea level. With an average elevation of about 0.8 kilometer (0.5 mile), continents lie close to sea level, except for limited areas of mountainous terrain. By contrast, the average depth of the ocean floor is about 3.8 kilometers (2.4 miles) below sea level.

The elevation difference between the continents and ocean basins is primarily the result of differences in their respective densities and thicknesses. Recall that the continents average 35 to 40 kilometers in thickness and are composed of granitic rocks having a density of about 2.7 g/cm^3. The basaltic rocks that comprise the oceanic crust average only 7 kilometers thick and have an average density of about 3.0 g/cm^3. Thus, the thicker and less dense continental crust is more buoyant than the oceanic crust. As a result, continental crust

*Liquid water has a density of 1 g/cm^3; therefore, the density of basalt is three times that of water

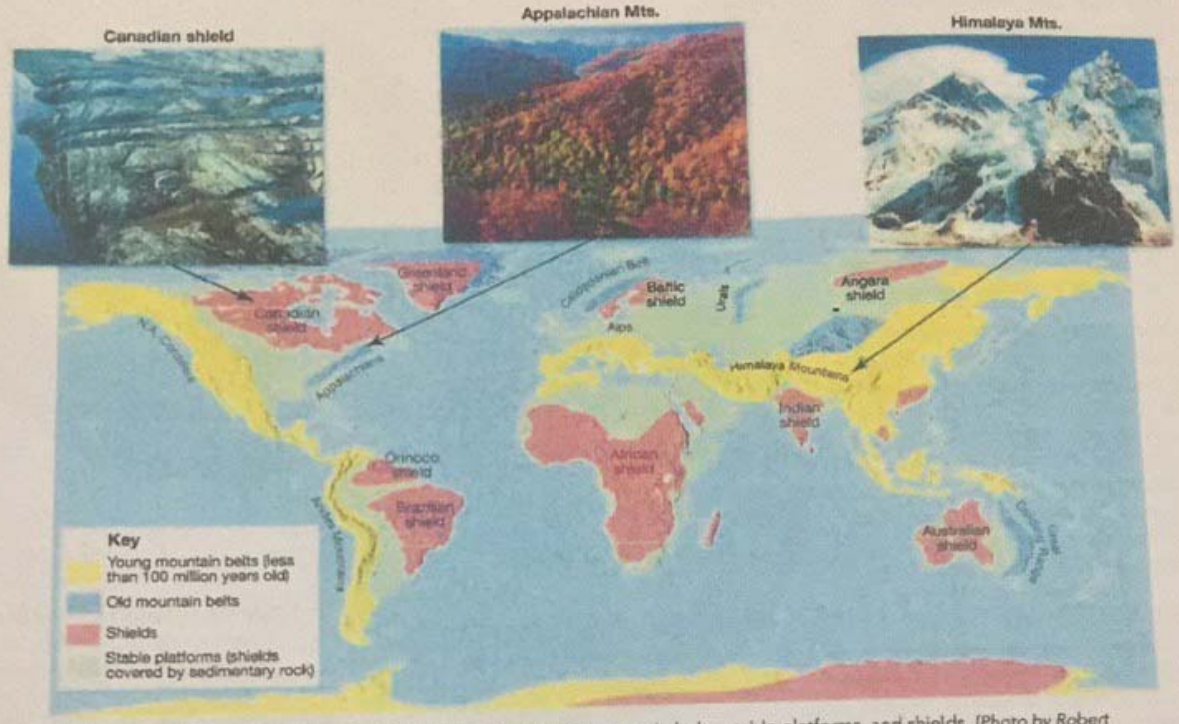


FIGURE 1.28 This map shows the general distribution of Earth's mountain belts, stable platforms, and shields. *[Photo by Robert Hildebrand Photography (left), CORBIS (middle), Images Source Pink/Alamy (right)]*

geologic past and will continue to shape it in the future.

CONCEPT CHECK 1.11

1. Describe the general distribution of Earth's youngest mountains.
2. What is the difference between shields and stable platforms?
3. What are the three major regions of the ocean floor and some features associated with each?

Dynamic Earth

Plate Tectonics: Introduction

Earth is a dynamic planet! If we could go back in time a few hundred million years, we would find the face of our planet dramatically different from what we see today. There would be no Mount St. Helens, Rocky Mountains, or Gulf of Mexico. Moreover, we would find continents having different sizes and shapes and located in different positions than today's landmasses. By contrast, over the past few billion years the Moon's surface has remained essentially unchanged—only a few craters have been added.

A Brief Introduction to the Theory of Plate Tectonics

During the past several decades a great deal has been learned about the workings of our dynamic planet. This span has seen an unequaled revolution in our understanding of Earth. The revolution had its beginnings in the early part of the 20th century with the radical proposal of *continental drift*—the idea that continents move about the face of the planet. This proposal contradicted the long established view that continents and ocean basins are permanent and stationary features on the face of Earth. For that reason, the notion of drifting continents was received with great skepticism and even ridicule. More than 50 years passed before enough data were gathered to transform this controversial hypothesis into a sound theory that wove together the basic processes known to operate on Earth. The theory that finally emerged, called the **theory of plate tectonics,** provided geologists with the first comprehensive model of Earth's internal workings.

This section serves as a brief introduction to plate tectonics. It provides background and perspective that will be helpful in some upcoming discussions, especially those dealing with volcanoes and igneous activity (Chapter 4) and metamorphic rocks (Chapter 7). Later Chapter 15, "Plate Tectonics: A Scientific Revolution Unfolds," explores the theory in detail.

According to the plate tectonics model, Earth's rigid outer shell (*lithosphere*) is broken into numerous slabs called **lithospheric plates** or simply **plates,** which are in

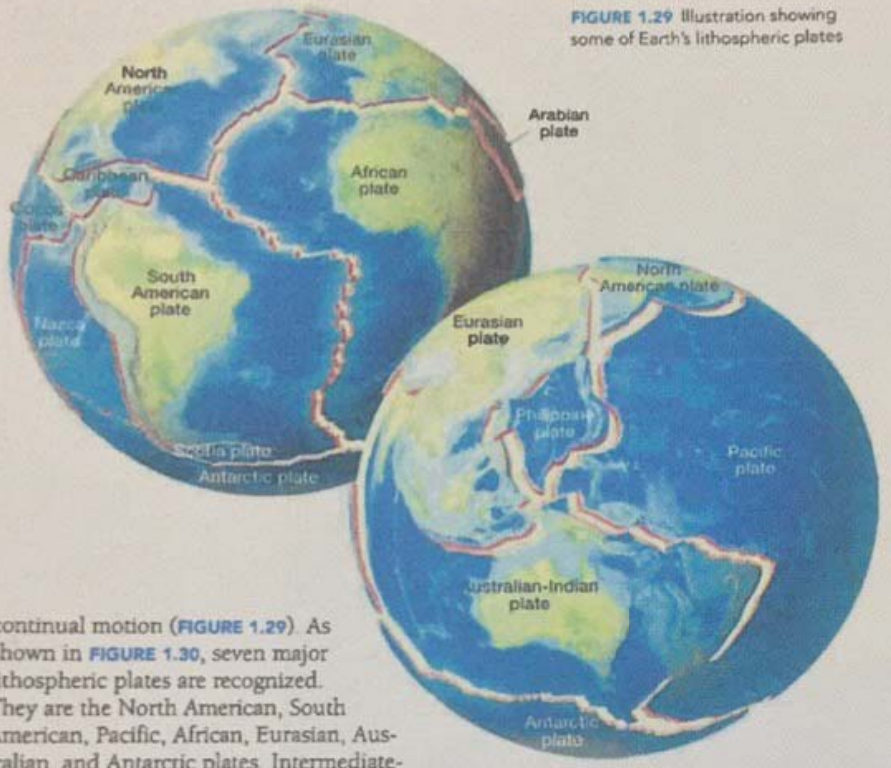


FIGURE 1.29 Illustration showing some of Earth's lithospheric plates

continual motion (FIGURE 1.29). As shown in FIGURE 1.30, seven major lithospheric plates are recognized. They are the North American, South American, Pacific, African, Eurasian, Australian, and Antarctic plates. Intermediate-size plates include the Caribbean, Nazca, Philippine, Arabian, Cocos, and Scotia plates. In addition, over a dozen smaller plates have been identified but are not shown in Figure 1.30. Note that several large plates include an entire continent plus a large area of seafloor (for example, the South American plate). However, none of the plates is defined entirely by the margins of a single continent.

The lithospheric plates move relative to each other at a very slow but continuous rate that averages about 5 centimeters (2 inches) a year. This movement is ultimately driven by the unequal distribution of heat within Earth. Hot material found deep in the mantle moves slowly upward and serves as one part of our planet's internal convective system. Concurrently, cooler, denser slabs of lithosphere descend back into the mantle, setting Earth's rigid outer shell in motion. Ultimately, the titanic, grinding movements of Earth's lithospheric plates generate earthquakes, create volcanoes, and deform large masses of rock into mountains.

Plate Boundaries

Lithospheric plates move as coherent units relative to all other plates. Although the interiors of plates may experience some deformation, all major interactions among individual plates (and therefore most deformation) occur along their boundaries. In fact, the first attempts to outline plate boundaries were made using locations of earthquakes. Later work showed that plates are bounded by three distinct types of boundaries, which are differentiated by the type of relative movement they exhibit. These boundaries are depicted at the bottom of Figure 1.30 and are briefly described here:

1. **Divergent boundaries**—where plates move apart, resulting in upwelling of material from the mantle to create new seafloor (Figure 1.30A).
2. **Convergent boundaries**—where plates move together, resulting in the subduction (consumption) of oceanic lithosphere into the mantle (Figure 1.30B). Convergence can also result in the collision of two continental margins to create a major mountain system.
3. **Transform fault boundaries**—where plates grind past each other without the production or destruction of lithosphere (Figure 1.30C).

If you examine Figure 1.30, you can see that each large plate is bounded by a combination of these boundaries. Movement along one boundary requires that adjustments be made at the others.

DIVERGENT BOUNDARIES. Plate spreading (divergence) occurs mainly along the oceanic ridge. As plates pull apart, the fractures created are immediately filled with molten rock that wells up from the asthenosphere below (FIGURE 1.31). This hot material slowly cools to become solid rock, producing new slivers of seafloor. This happens again and again over millions of years, adding thousands of square kilometers of new seafloor.

This mechanism has created the floor of the Atlantic Ocean during the past 160 million years and is appropriately called **seafloor spreading** (Figure 1.31). Because seafloor spreading is the dominant process associated with divergent boundaries, these zones are sometimes referred to as *spreading centers*. The rate of seafloor spreading varies considerably from one spreading center to another. Spreading rates of only 2.5 centimeters (1 inch) per year are typical in the North Atlantic, whereas much faster rates (20 centimeters, 8 inches per year) have been measured along the East Pacific Rise. Even the most rapid rates of spreading are slow on the scale of human history. Nevertheless, the slowest rate of lithosphere production is rapid enough to have created all of Earth's ocean basins over the last 200 million years. In fact, none of the ocean floor that has been dated exceeds 180 million years in age.

Along divergent boundaries where molten rock emerges, the oceanic lithosphere is elevated, because it is hot and occupies more volume than do cooler rocks. Worldwide, this elevated zone (the oceanic ridge) extends for over 70,000 kilometers (43,000 miles) through all major ocean basins (Figure 1.27). As new lithosphere is formed along the oceanic

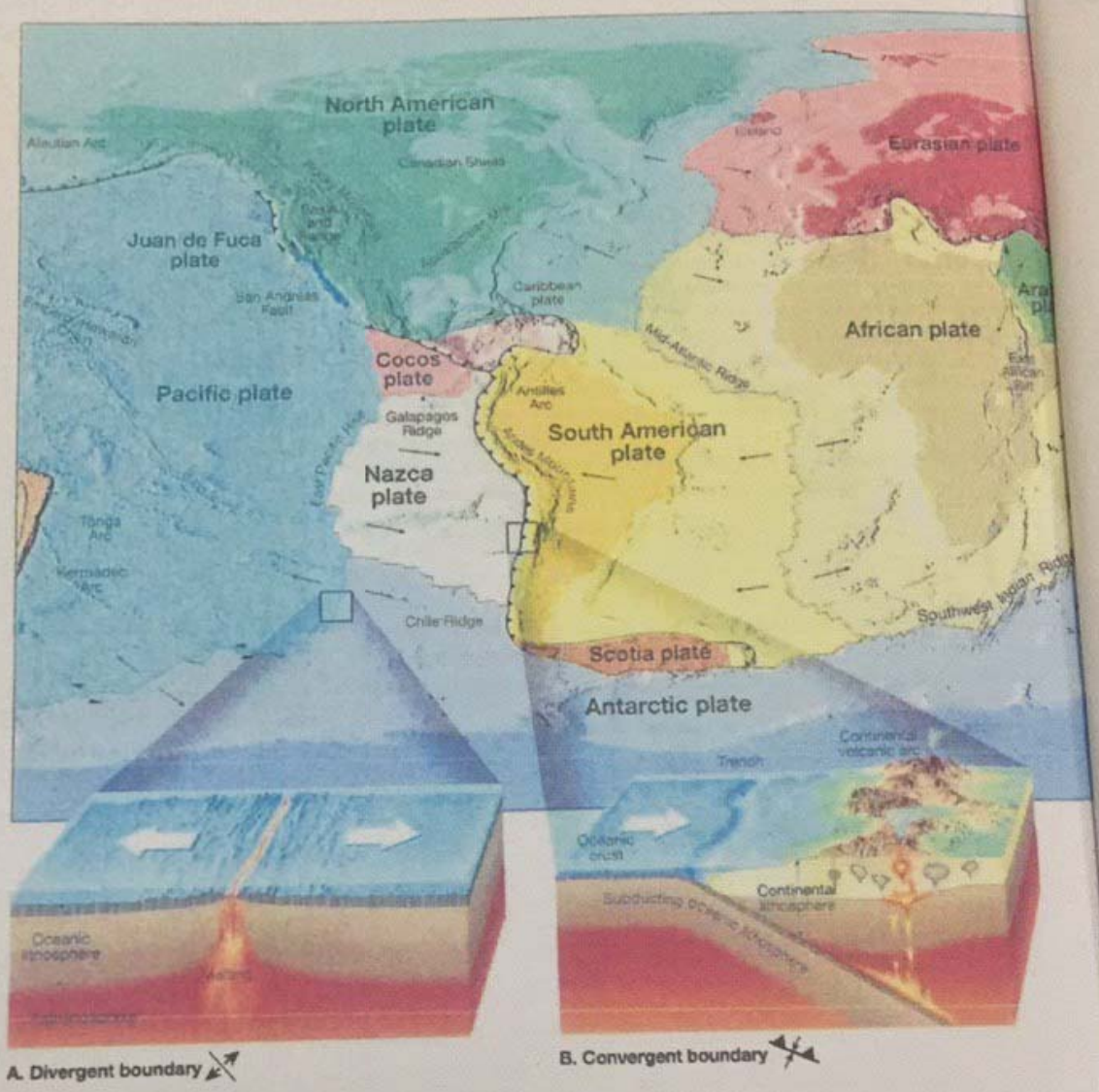


FIGURE 1.30 Mosaic of rigid plates that constitute Earth's outer shell. (After W. B. Hamilton, U.S. Geological Survey)

ridge, it is slowly yet continually displaced from the zone of upwelling along the ridge axis. Thus, it begins to cool and contract, thereby increasing in density. This thermal contraction accounts for the greater ocean depths that exist away from the ridge. In addition, cooling causes the mantle rocks below the oceanic crust to strengthen, thereby adding to the plate's thickness. Stated another way, the thickness of oceanic lithosphere is age dependent. The older (cooler) it is, the greater its thickness.

CONVERGENT BOUNDARIES. Although new lithosphere is constantly being added at the oceanic ridges, the planet is not growing in size—its total surface area remains constant. To accommodate the newly created lithosphere, older oceanic plates return to the mantle along *convergent boundaries*. As two plates slowly converge, the leading edge of one slab is bent downward, allowing it to slide beneath the other. The surface expression produced by the descending plate is a *deep-ocean trench*, like the Peru–Chile trench illustrated in Figures 1.27, 1.30, and 1.31.

Plate margins where oceanic crust is being consumed are called **subduction zones.** Here, as the subducted plate moves downward, it enters a high-temperature, high-pressure environment. Some subducted materials, as well as more voluminous amounts of the asthenosphere located above the subducting slab, melt and migrate upward into the overriding plate. Occasionally this molten rock may reach

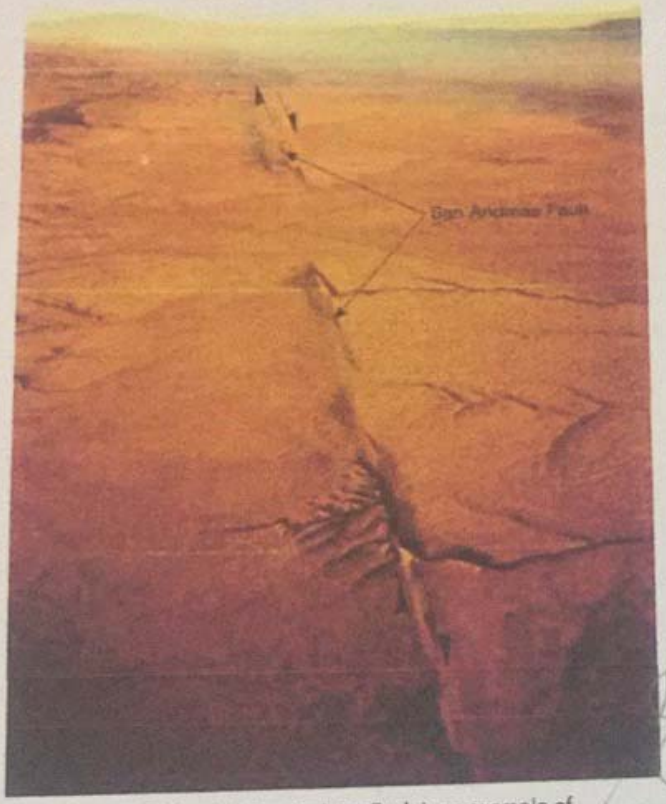


FIGURE 1.34 California's San Andreas Fault is an example of a transform fault boundary *(Photo by Michael Collier)*

New plate boundaries are created in response to changes in the forces acting on the lithosphere. For example, a relatively new divergent boundary is located in Africa, in a region known as the East African Rift Valleys. If spreading continues in this region, the African plate will split into two plates, separated by a new ocean basin. At other locations plates carrying continental crust are moving toward each other. Eventually these continents may collide and be sutured together. Thus, the boundary that once separated these plates disappears, and two plates become one.

DID YOU KNOW?
The average rate at which plates move relative to each other is roughly the same rate at which human fingernails grow.

As long as temperatures within the interior of our planet remain significantly higher than those at the surface, material within Earth will continue to circulate. This internal flow, in turn, will keep the rigid outer shell of Earth in motion. Thus, while Earth's internal heat engine is operating, the positions and shapes of the continents and ocean basins will change and Earth will remain a dynamic planet.

In the remaining chapters we will examine in more detail the workings of our dynamic planet in light of the plate tectonics theory.

CONCEPT CHECK 1.12

1. What are the three basic types of plate boundaries and the relative movement associated with each?
2. Name the type of plate boundary with which each of the following is associated: deep-ocean trench, seafloor spreading, San Andreas Fault, subduction, oceanic ridge, and Himalaya Mountains.

CHAPTER ONE

An Introduction to Geology in Review

Geology means "the study of Earth." The two broad areas of the science of geology are (1) *physical geology*, which examines the materials composing Earth and the processes that operate beneath and upon its surface; and (2) *historical geology*, which seeks to understand the origin of Earth and its development through time.

The relationship between people and the natural environment is an important focus of geology. This includes natural hazards, resources, and human influences on geologic processes.

During the 17th and 18th centuries, *catastrophism* influenced the formulation of explanations about Earth. Catastrophism states that Earth's landscapes developed over short time spans primarily as a result of great catastrophes. By contrast, *uniformitarianism*, one of the fundamental principles of modern geology advanced by James Hutton in the late 1700s, states that the physical, chemical, and biological laws that operate today have also operated in the geologic past. The idea is often summarized as "*The present is the key to the past.*" Hutton argued that processes that appear to be slow acting could, over long spans of time, produce effects that are just as great as those resulting from sudden catastrophic events.

Using the principles of relative dating, the placing of events in their proper sequence or order without knowing their age in years, scientists developed a geologic time scale during the 19th

century. Relative dates can be established by applying such principles as the *law of superposition* and the *principle of fossil succession*.

⊙ All science is based on the assumption that the natural world behaves in a consistent and predictable manner. The process by which scientists gather facts and formulate scientific *hypotheses* and *theories* is called the *scientific method*. To determine what is occurring in the natural world, scientists often (1) collect facts, (2) ask questions and develop hypotheses that may answer these questions, (3) develop observations and experiments to test the hypotheses, and (4) accept, modify, or reject hypotheses on the basis of extensive testing. Other discoveries represent purely theoretical ideas that have stood up to extensive examination. Still other scientific advancements have been made when a totally unexpected happening occurred during an experiment.

⊙ Earth's physical environment is traditionally divided into three major parts: the solid Earth, or *geosphere*; the water portion of our planet, the *hydrosphere*; and Earth's gaseous envelope, the *atmosphere*. In addition, the *biosphere*, the totality of life on Earth, interacts with each of the three physical realms and is an equally integral part of Earth.

⊙ Although each of Earth's four spheres can be studied separately, they are all related in a complex and continuously interacting whole that we call the *Earth system*. *Earth system science* uses an interdisciplinary approach to integrate the knowledge of several academic fields in the study of our planet and its global environmental problems.

⊙ A *system* is a group of interacting parts that form a complex whole. *Closed systems* are those in which energy moves freely in and out, but matter does not enter or leave the system. In an open system, both energy and matter flow into and out of the system.

⊙ Most natural systems have mechanisms that tend to enhance change, called *positive feedback mechanisms*, and other mechanisms, called *negative feedback mechanisms*, that tend to resist change and thus stabilize the system.

⊙ The two sources of energy that power the Earth system are (1) the Sun, which drives the external processes that occur in the atmosphere, hydrosphere, and at Earth's surface, and (2) heat from Earth's interior that powers the internal processes that produce volcanoes, earthquakes, and mountains.

⊙ The *rock cycle* is one of the many cycles or loops of the Earth system in which matter is recycled. The rock cycle is a means of viewing many of the interrelationships of geology. It illustrates the origin of the three basic rock groups and the role of various geologic processes in transforming one rock type into another.

⊙ The *nebular* theory describes the formation of the solar system. The planets and Sun began forming about 5 billion years ago from a large cloud of dust and gases. As the cloud contracted, it began to rotate and assume a disk shape. Material that was gravitationally pulled toward the center became the *protosun*. Within the rotating disk, small centers, called *planetesimals*, swept up more and more of the cloud's debris. Because of the high temperatures near the Sun, the inner planets were unable to accumulate many of the elements that vaporize at low temperatures. Because of the very cold temperatures existing far from the Sun, the large outer planets consist of huge amounts of ices and lighter materials. These substances account for the comparatively large sizes and low densities of the outer planets.

⊙ Earth's internal structure is divided into layers based on differences in chemical composition and on the basis of changes in physical properties. Compositionally, Earth is divided into a thin outer *crust*, a solid rocky *mantle*, and a dense *core*. Other layers, based on physical properties, include the lithosphere, asthenosphere, lower mantle, outer core, and inner core.

⊙ Two principal divisions of Earth's surface are the *continents* and *ocean basins*. A significant difference is their relative levels. The elevation differences between continents and ocean basins is primarily the result of differences in their respective densities and thicknesses.

⊙ The largest features of the continents can be divided into two categories: *mountain belts* and the *stable interior*. The ocean floor is divided into three major topographic units: *continental margins*, *deep-ocean basins*, and *oceanic (mid-ocean) ridges*.

⊙ The *theory of plate tectonics* provides a comprehensive model of Earth's internal workings. It holds that Earth's rigid outer lithosphere consists of several segments called *lithospheric plates* that are slowly and continually in motion relative to one another. Most earthquakes, volcanic activity, and mountain building are associated with the movements of these plates.

⊙ The three distinct types of plate boundaries are (1) *divergent boundaries*, where plates move apart; (2) *convergent boundaries*, where plates move together, causing one to go beneath another, or where plates collide, which occurs when the leading edges are made of continental crust; and (3) *transform fault boundaries*, where plates slide past one another.

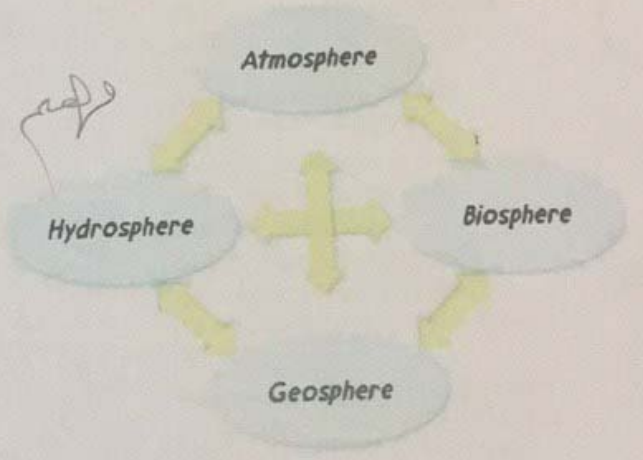
Key Terms

abyssal plains (p. 26)
asthenosphere (p. 24)
atmosphere (p. 13)
biosphere (p. 14)
catastrophism (p. 4)
closed systems (p. 15)
continental margin (p. 25)
continental rise (p. 26)
continental shelf (p. 25)
continental slope (p. 25)
convergent boundaries (p. 29)
core (p. 24)
crust (p. 23)
deep-ocean basins (p. 26)
deep-ocean trenches (p. 26)
divergent boundaries (p. 29)
Earth system science (p. 15)
fossils (p. 7)
fossil succession, principle of (p. 7)
geology (p. 2)
geosphere (p. 14)
historical geology (p. 2)
hydrosphere (p. 13)
hypothesis (p. 8)
igneous rocks (p. 20)
inner core (p. 24)
interface (p. 16)
lithosphere (p. 24)
lithospheric plates (p. 28)
lower mantle (p. 24)
magma (p. 20)

mantle (p. 24)
metamorphic rock (p. 20)
nebular theory (p. 20)
negative feedback mechanism (p. 16)
oceanic (mid-ocean) ridge (p. 27)
open systems (p. 16)
outer core (p. 24)
physical geology (p. 2)
plates (p. 28)
plate tectonics, theory of (p. 28)
positive feedback mechanism (p. 16)
relative dating (p. 7)
rock cycle (p. 18)
seafloor spreading (p. 29)
seamounts (p. 26)
sediment (p. 20)
sedimentary rock (p. 20)
shields (p. 25)
solar nebula (p. 20)
stable platforms (p. 25)
subduction zone (p. 30)
superposition, law of (p. 7)
system (p. 15)
theory (p. 9)
transform fault boundaries (p. 29)
uniformitarianism (p. 5)

GIVE IT SOME THOUGHT

1. Refer to Figure 1.17. Which of the four main components of the Earth System (atmosphere, biosphere, geosphere, hydrosphere) were involved in the natural disaster at Caraballeda, Venezuela? Describe how each of the components you list contributed to the event depicted in the photo.
2. Earth is 4.6 billion years old. This number does not come from studying Earth rocks since Earth's oldest rocks only date to about 4 billion years. The number comes from studying lunar rocks and meteorites. Based on what you know about the formation of our solar system, how old do you think the other planets are? Why did you assign this age or these ages?
3. Why do geologists look to meteorites to tell us about the internal structure of Earth?
4. After entering a dark room, you turn on a wall switch but the light does not come on. Suggest at least three hypotheses that might explain this observation.
5. Each of the following statements may either be a hypothesis (H), a theory (T), or an observation (O). Use one of these letters to identify each statement. Briefly explain each choice.
   a. A scientist proposes that a recently discovered ring-shaped structure is the remains of an ancient meteorite crater.
   b. The Redwall Formation in the Grand Canyon is composed primarily of limestone.
   c. Earth is composed of several large plates that move and interact with each other.
   d. Since 1885, the terminus of Canada's Athabasca Glacier has receded 1.5 kilometers.
6. Consider the possible results of the following scenario and describe one positive and one negative feedback. *Earth is getting warmer, consequently evaporation is increasing.*
7. Refer to Figure 1.22. How does the rock cycle diagram, in particular the process arrows, support the fact that sedimentary rocks are the most abundant rock type on the surface of Earth?
8. Look at the concept map linking the four spheres of the Earth system. All of the spheres are linked by arrows representing processes by which the spheres interact and influence each other. For each arrow, describe at least one process.



9. Refer to Figure 1.27 showing major features and topography of Earth's solid surface. Give an example of where you would expect to find (a) very old, deformed rocks, and (b) young undisturbed rocks. Explain your choices.

Companion Website

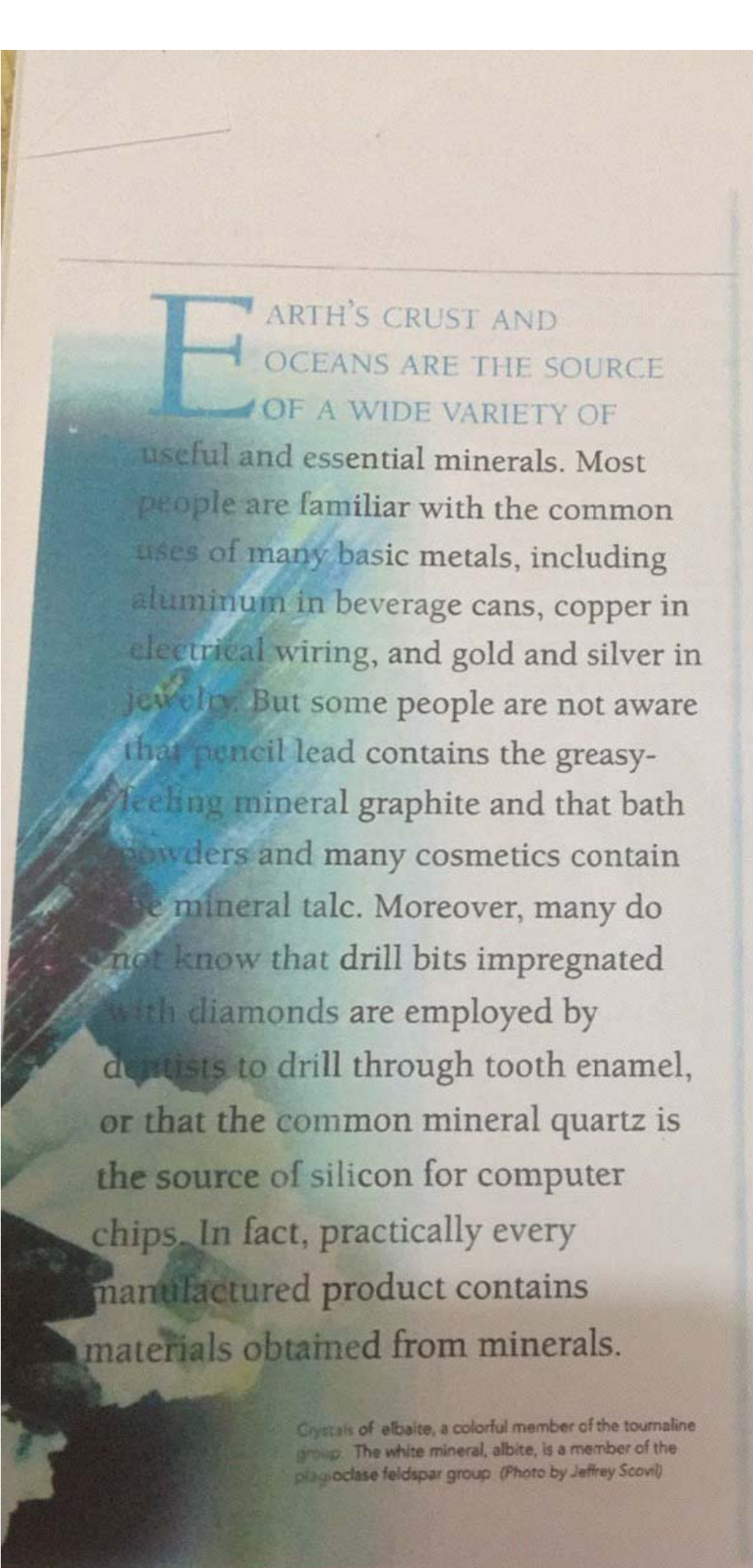
mygeoscienceplace

www.mygeoscienceplace.com

The *Essentials of Geology, 11e* companion Website contains numerous multimedia resources accompanied by assessments to aid in your study of the topics in this chapter. The use of this site's learning tools will help improve your understanding of geology. Utilizing the access code that accompanies this text, visit **www.mygeoscienceplace.com** in order to:

- **Review** key chapter concepts.
- **Read** with links to the Pearson eText and to chapter-specific web resources.
- **Visualize** and comprehend challenging topics using the learning activities in *GEODe: Essentials of Geology* and the *Geoscience Animations Library*.
- **Test** yourself with online quizzes.

EARTH'S CRUST AND OCEANS ARE THE SOURCE OF A WIDE VARIETY OF useful and essential minerals. Most people are familiar with the common uses of many basic metals, including aluminum in beverage cans, copper in electrical wiring, and gold and silver in jewelry. But some people are not aware that pencil lead contains the greasy-feeling mineral graphite and that bath powders and many cosmetics contain the mineral talc. Moreover, many do not know that drill bits impregnated with diamonds are employed by dentists to drill through tooth enamel, or that the common mineral quartz is the source of silicon for computer chips. In fact, practically every manufactured product contains materials obtained from minerals.



Crystals of elbaite, a colorful member of the tourmaline group. The white mineral, albite, is a member of the plagioclase feldspar group. *(Photo by Jeffrey Scovil)*

FOCUS ON CONCEPTS

To assist you in learning the important concepts in this chapter, focus on the following questions:

- What are minerals, and how are they different from rocks?
- What are the smallest particles of matter?
- How do atoms bond?
- How do isotopes of the same element vary, and why are some isotopes radioactive?
- What are some of the physical and chemical properties of minerals? How can these properties distinguish one mineral from another?
- What are the eight elements that make up most of Earth's continental crust?
- What is the most abundant mineral group? What do all minerals within this group have in common?
- When is the term ore used with reference to a mineral? What are the common ores of iron and lead?

Minerals: Building Blocks of Rocks

Matter and Minerals

ESSENTIALS OF GEOLOGY **Introduction**

We begin our discussion of Earth materials with an overview of **mineralogy** (*mineral* = mineral, *ology* = the study of), because minerals are the building blocks of rocks. In addition, minerals have been employed by humans for both useful and decorative purposes for thousands of years (FIGURE 2.1). The first minerals mined were flint and chert, which people fashioned into weapons and cutting tools. As early as 3700 BC, Egyptians began mining gold, silver, and copper; and by 2200 BC humans discovered how to combine copper with tin to make bronze, a strong, hard alloy. Later, humans developed a process to extract iron from minerals such as hematite—a discovery that marked the decline of the Bronze Age. By about 800 BC, iron-working technology had advanced to the point that weapons and many everyday objects were made of iron rather than copper, bronze, or wood. During the Middle Ages, mining of a variety of minerals was common throughout Europe, and the impetus for the formal study of minerals was in place.

The term *mineral* is used in several different ways. For example, those concerned with health and fitness extol the benefits of vitamins and minerals. The mining industry typically uses the word when referring to anything taken out of the ground, such as coal, iron ore, or sand and gravel. The guessing game known as *Twenty Questions* usually begins with the question, *Is it animal, vegetable, or mineral?* What criteria do geologists use to determine whether something is a mineral?

Geologists define **mineral** as *any naturally occurring inorganic solid that possesses an orderly crystalline structure and can be represented by a chemical formula.* Thus, Earth materials that are classified as minerals exhibit the following characteristics:

1. **Naturally occurring.** Minerals form by natural, geologic processes. Synthetic materials, meaning those produced in a laboratory or by human intervention, are not considered minerals.
2. **Solid substance.** Only solid crystalline substances are considered minerals. Ice (frozen water) fits this criterion and is considered a mineral, whereas liquid water and water vapor do not. The exception is mercury, which is found in its liquid form in nature.

FIGURE 2.1 Collection of well-developed quartz crystals found near Hot Springs, Arkansas. (Photo by Jeff Scovil)

3. **Orderly crystalline structure.** Minerals are crystalline substances, which means their atoms are arranged in an orderly, repetitive manner (FIGURE 2.2) This orderly packing of atoms is reflected in the regularly shaped objects called crystals. Some naturally occurring solids, such as volcanic glass (obsidian), lack a repetitive atomic structure and are not considered minerals.
4. **Generally inorganic.** Inorganic crystalline solids, such as ordinary table salt (halite), that are found naturally in the ground are considered minerals. (Organic compounds, on the other hand, are generally not. Sugar, a crystalline solid like salt but which comes from sugarcane or sugar beets, is a common example of such an organic compound.) Many marine animals secrete inorganic compounds, such as calcium carbonate (calcite), in the form of shells and coral reefs. If these materials are buried and become part of the rock record, they are considered minerals by geologists.
5. **Can be represented by a chemical formula.** Most minerals are chemical compounds having compositions that can be expressed by a chemical formula. For example, the common mineral quartz has the formula SiO_2, which indicates that quartz consists of silicon (Si) and oxygen (O) atoms in a ratio of one-to-two. This proportion of silicon to oxygen is true for any sample of pure quartz, regardless of its origin. However, the compositions of some minerals vary *within specific, well-defined limits*. This occurs because certain elements can substitute for others of similar size without changing the mineral's internal structure. An example is the mineral olivine in which either the element magnesium (Mg) or the element iron (Fe) may occupy the same site in the crystal structure. Therefore, olivine's formula, $(Mg, Fe)_2SiO_4$, expresses variability in the relative amounts of magnesium and iron. However, the ratio of magnesium plus iron (Mg + Fe) to silicon (Si) and oxygen (O) remains fixed at 2:1:4.

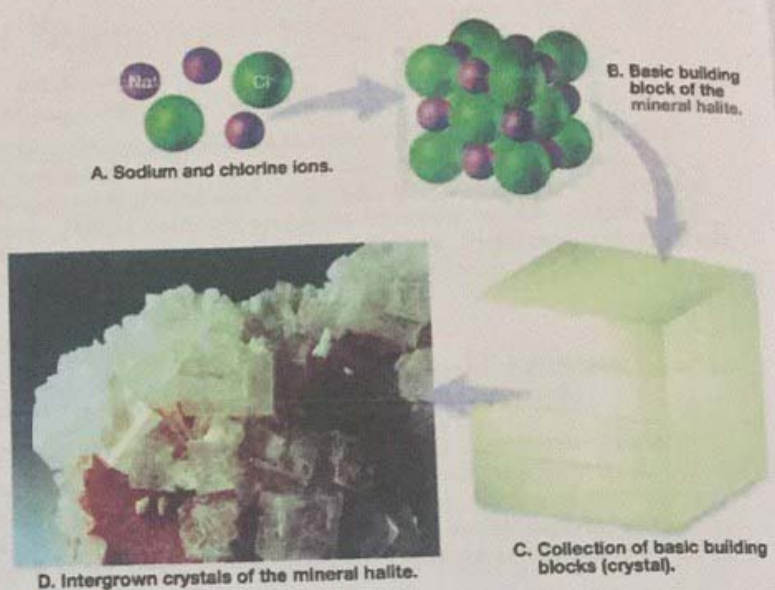


FIGURE 2.2 This diagram illustrates the orderly arrangement of sodium and chloride ions in the mineral halite. The arrangement of atoms into basic building blocks having a cubic shape results in regularly shaped cubic crystals. (Photo by Dennis Tasa)

In contrast to minerals, rocks are more loosely defined. Simply, a **rock** is any solid mass of mineral, or mineral-like, matter that occurs naturally as part of our planet. Most rocks, like the common rock granite shown in FIGURE 2.3, occur as aggregates of several different minerals. The term *aggregate* implies that the minerals are joined in such a way that their individual properties are retained. Note that the mineral constituents of granite can be easily identified. However, some rocks are composed almost entirely of one mineral. A common example is the sedimentary rock *limestone*, which consists of impure masses of the mineral calcite.

Some rocks are composed of nonmineral matter. These include the volcanic rocks *obsidian* and *pumice*, which are noncrystalline glassy substances, and *coal*, which consists of solid organic debris.

Although this chapter deals primarily with the nature of minerals, keep in mind that most rocks are simply aggregates of

FIGURE 2.3 Most rocks are aggregates of two or more minerals. Shown here is a hand sample of the igneous rock granite and three of its major constituent minerals. (Photos by E. J. Tarbuck)

DID YOU KNOW?

Archaeological finds show that more than 2000 years ago the Romans used lead for pipes to transport water within their buildings. In fact, Roman smelting of lead and copper ores between 500 BC and AD 300 caused a small but significant rise in atmospheric pollution, as recorded in Greenland ice cores.

Isotopes and Radioactive Decay

The **mass number** of an atom is simply the total number of its protons and neutrons. All atoms of a particular element have the same number of protons but they may have varying numbers of neutrons. Atoms with the same number of protons but different numbers of neutrons are **isotopes** of that element. Isotopes of the same element are labeled by placing the mass number after the element's name or symbol. For example, carbon has three well-known isotopes. One has a mass number of 12 (carbon-12), another has a mass number of 13 (carbon-13), and the third, carbon-14, has a mass number of 14. Carbon-12 must also have six neutrons to give it a mass number of 12. Carbon-14, on the other hand, has six protons plus eight neutrons to give it a mass number of 14.

In chemical behavior, all isotopes of the same element are nearly identical. To distinguish among them is like trying to differentiate identical twins, with one weighing slightly more than the other. Because isotopes of the same element exhibit the same chemical behavior, they often become parts of the same mineral. For example, when the mineral calcite ($CaCO_3$) forms, some of its carbon atoms are carbon-12, and some are carbon-14.

The nuclei of most atoms are stable. However, many elements do have isotopes in which the nuclei are unstable—carbon-14 is one example of an unstable isotope. In this context, *unstable* means that the nuclei change through a random process called **radioactive decay.** During radioactive decay, unstable isotopes radiate energy and emit particles. The rates at which unstable isotopes decay are measurable. Therefore, certain radioactive atoms are used to determine the ages of fossils, rocks, and minerals. A discussion of radioactive decay and its applications in dating past geologic events appears in Chapter 18.

CONCEPT CHECK 2.4

● What is an isotope?

Physical Properties of Minerals

Matter and Minerals

ESSENTIALS OF GEOLOGY **Physical Properties of Minerals**

Minerals have definite crystalline structures and chemical compositions that give them unique sets of physical and chemical properties shared by all samples of that mineral. For example, all specimens of halite have the same hardness, the same density, and break in a similar manner. Because a mineral's internal structure and chemical composition are difficult to determine without the aid of sophisticated tests and equipment, the more easily recognized physical properties are frequently used in identification.

Optical Properties

Of the many optical properties of minerals, their luster, their ability to transmit light, their color, and their streak are most frequently used for mineral identification.

LUSTER. The appearance or quality of light reflected from the surface of a mineral is known as **luster.** Minerals that have the appearance of metals, regardless of color, are said to have a *metallic luster* (FIGURE 2.9). Some metallic minerals, such as native copper and galena, develop a dull coating or tarnish when exposed to the atmosphere. Because they are not as shiny as samples with freshly broken surfaces, these samples are often said to exhibit a *submetallic luster.*

Most minerals have a *nonmetallic luster* and are described using various adjectives such as *vitreous* or *glassy.* Other nonmetallic minerals are described as having a *dull* or *earthy luster* (a dull appearance like soil) or a *pearly luster* (such as a pearl or the inside of a clamshell). Still others exhibit a *silky luster* (like satin cloth) or a *greasy luster* (as though coated in oil).

THE ABILITY TO TRANSMIT LIGHT. Another optical property used in the identification of minerals is the ability to transmit light. When no light is transmitted, the mineral is described as *opaque;* when light, but not an image, is transmitted through a mineral it is said to be *translucent.* When both light and an image are visible through the sample, the mineral is described as *transparent.*

COLOR. Although **color** is generally the most conspicuous characteristic of any mineral, it is considered a diagnostic property of only a few minerals. Slight impurities in the common mineral quartz, for example, give it a variety of tints including pink, purple, yellow, white, gray, and even black (FIGURE 2.10). Other minerals, such as tourmaline, also exhibit a variety of hues, with multiple colors sometimes occurring in the same sample. Thus, the use of color as a means of identification is often ambiguous or even misleading.

FIGURE 2.9 The freshly broken sample of galena (right) displays a metallic luster, while the sample on the left is tarnished and has a submetallic luster. (Photo courtesy of E. J. Tarbuck)

DID YOU KNOW?
The name crystal is derived from the Greek (krystallos, meaning "ice") and was applied to quartz crystals. The ancient Greeks thought quartz was water that had crystallized at high pressures deep inside Earth.

FIGURE 2.10 Quartz. Some minerals, such as quartz, occur in a variety of colors. These samples include crystal quartz (colorless), amethyst (purple quartz), citrine (yellow quartz), and smoky quartz (gray to black). *(Photo courtesy of E. J. Tarbuck)*

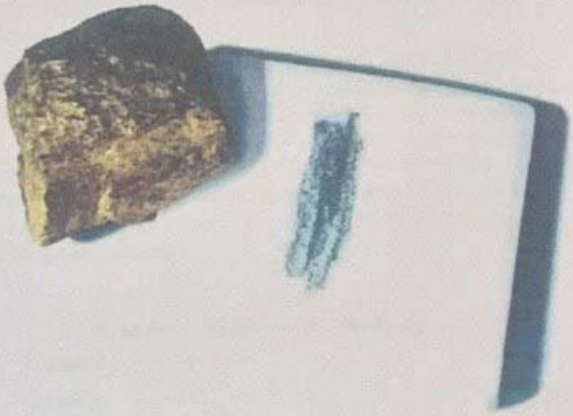


FIGURE 2.11 Although the color of a mineral is not always helpful in identification, the streak, which is the color of the powdered mineral, can be very useful. *(Photo by Dennis Tasa)*

STREAK. The color of the mineral in powdered form, called **streak,** is often useful in identification. A mineral's streak is obtained by rubbing it across a *streak plate* (a piece of unglazed porcelain) and observing the color of the mark it leaves (FIGURE 2.11). Although the color of a mineral may vary from sample to sample, its streak is usually consistent in color.

Streak can also help distinguish between minerals with metallic luster and those with nonmetallic luster. Metallic minerals generally have a dense, dark streak, whereas minerals with nonmetallic luster typically have a light colored streak.

It should be noted that not all minerals produce a streak when rubbed across a streak plate. For example, the mineral quartz is harder than a porcelain streak plate. Therefore, no streak is observed using this method.

Crystal Shape or Habit

Mineralogists use the term **crystal shape** or **habit** to refer to the common or characteristic shape of a crystal or aggregate of crystals. A few minerals exhibit somewhat regular polygons that are helpful in their identification. For example, magnetite crystals sometimes occur as octahedrons, garnets often form dodecahedrons, and halite and fluorite crystals tend to grow as cubes or near cubes. While most minerals have only one common habit, a few have two or more characteristic crystal shapes such as the pyrite sample shown in FIGURE 2.12.

Aragonite crystals *(Photo by Ross Frid)*

minerals

FIGURE 2.12 Although most minerals exhibit only one common crystal shape, some, such as pyrite, have two or more characteristic habits. *(Photos by Dennis Tasa)*

By contrast, some minerals rarely develop perfect geometric forms. Many of these, however, develop other characteristic shapes useful for identification. Some minerals tend to grow equally in all three dimensions, whereas others tend to be elongated in one direction, or flattened if growth in one dimension is suppressed. Commonly used terms to describe these and other crystal habits include *equant* (equidimensional), *bladed*, *fibrous*, *tabular*, *prismatic*, *platy*, *blocky*, and *botryoidal*. Some of these habits are pictured in **FIGURE 2.13**.

Mineral Strength

How easily minerals break or deform under stress is determined by the type and strength of the chemical bonds that hold the crystals together. Mineralogists use terms including *tenacity*, *hardness*, *cleavage*, and *fracture* to describe mineral strength and how minerals break when stress is applied.

TENACITY. The term **tenacity** describes a mineral's toughness, or its resistance to breaking or deforming. Minerals that are ionically bonded, such as fluorite and halite, tend to be *brittle* and shatter into small pieces when struck. By contrast, minerals with metallic bonds, such as native copper, are *malleable*, or easily hammered into different shapes. Minerals, including gypsum and talc, that can be cut into thin shavings are described as *sectile*. Still others, notably the micas, are *elastic* and will bend and snap back to their original shape after the stress is released.

HARDNESS. One of the most useful diagnostic properties is **hardness**, a measure of the resistance of a mineral to abrasion or scratching. This property is determined by rubbing a mineral of unknown hardness against one of known hardness, or vice versa. A numerical value of hardness can by obtained by using the **Mohs scale** of hardness, which consists of 10 minerals arranged in order from 1 (softest) to 10 (hardest), as shown in **FIGURE 2.14A**. It should be noted that the Mohs scale is a relative ranking, and it does not imply that mineral number 2, gypsum, is twice as hard as mineral 1, talc. In fact, gypsum is only slightly harder than talc, as **FIGURE 2.14B** indicates.

In the laboratory, other common objects can be used to determine the hardness of a mineral. These include a human fingernail, which has a hardness of about 2.5, a copper penny (3.5), and a piece of glass (5.5). The mineral gypsum, which has a

FIGURE 2.13 Some common crystal habits. **A. *Bladed.*** Elongated crystals that are flattened in one direction. **B. *Prismatic.*** Elongated crystals with faces that are parallel to a common direction. **C. *Banded.*** Minerals that have stripes or bands of different color or texture. **D. *Botryoidal.*** Groups of intergrown crystals resembling a bunch of grapes. *(Photos by Dennis Tasa)*

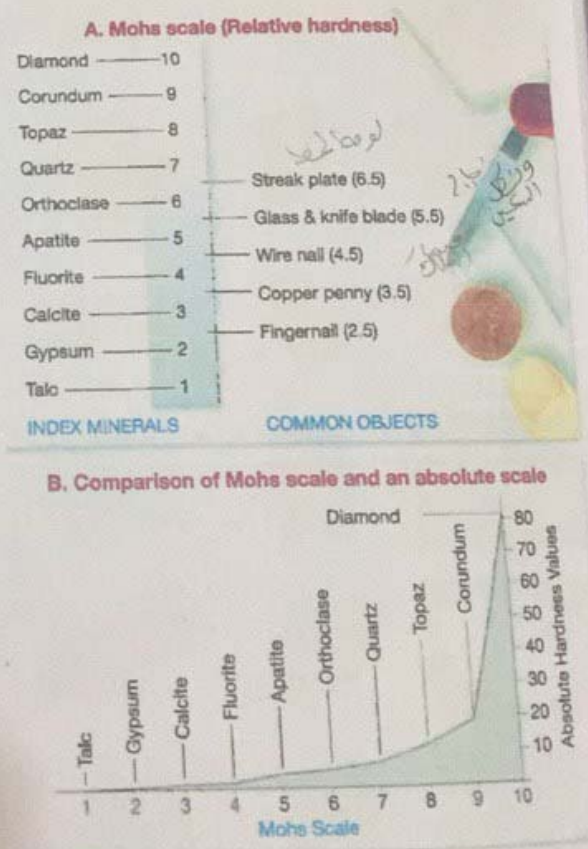


FIGURE 2.14 Hardness scales. **A.** Mohs scale of hardness, with the hardness of some common objects. **B.** Relationship between Mohs relative hardness scale and an absolute hardness scale.

hardness of 2, can be easily scratched with a fingernail. On the other hand, the mineral calcite, which has a hardness of 3, will scratch a fingernail but will not scratch glass. Quartz, one of the hardest common minerals, will easily scratch glass. Diamonds, hardest of all, scratch anything, including other diamonds.

FIGURE 2.15 The thin sheets shown here were produced by splitting a mica (muscovite) crystal parallel to its perfect cleavage *(Photo by Chip Clark)*

CLEAVAGE. In the crystal structure of many minerals, some atomic bonds are weaker than others. It is along these weak bonds that minerals tend to break when they are stressed. **Cleavage** (*Kleiben* = carve) is the tendency of a mineral to break (cleave) along planes of weak bonding. Not all minerals have cleavage, but those that do can be identified by the relatively smooth, flat surfaces that are produced when the mineral is broken.

The simplest type of cleavage is exhibited by the micas (**FIGURE 2.15**). Because these minerals have very weak bonds in one direction, they cleave to form thin, flat sheets. Some minerals have excellent cleavage in one, two, three, or more directions, whereas others exhibit fair or poor cleavage, and still others have no cleavage at all. When minerals break evenly in more than one direction, cleavage is described by *the number of cleavage directions and the angle(s) at which they meet* (**FIGURE 2.16**).

Each cleavage surface that has a different orientation is counted as a different direction of cleavage. For example, some minerals cleave to form six-sided cubes. Because cubes are defined by three different sets of parallel planes that intersect at 90-degree angles, cleavage is described as *three directions of cleavage that meet at 90 degrees.*

Do not confuse cleavage with crystal shape. When a mineral exhibits cleavage, it will break into pieces that all have the same geometry. By contrast, the smooth-sided quartz crystals shown in Figure 2.1 (p. 38) do not have cleavage. If broken, they fracture into shapes that do not resemble one another or the original crystals.

FRACTURE. Minerals having chemical bonds that are equally, or nearly equally, strong in all directions exhibit a property called **fracture**. When minerals fracture, most produce uneven surfaces and are described as exhibiting *irregular fracture*. However, some minerals, such as

Number of Cleavage Directions	Shape	Sketch	Directions of Cleavage	Sample
1	Flat sheets			Muscovite
2 at 90°	Elongated form with rectangle cross section (prism)			Feldspar
2 not at 90°	Elongated form with parallelogram cross section (prism)			Hornblende
3 at 90°	Cube			Halite
3 not at 90°	Rhombohedron			Calcite
4	Octahedron			Fluorite

FIGURE 2.16 Common cleavage directions exhibited by minerals *(Photos by E. J. Tarbuck and Dennis Tasa)*

FIGURE 2.17 Conchoidal fracture The smooth curved surfaces result when minerals break in a glasslike manner. *(Photo by E. J. Tarbuck)*

quartz, break into smooth, curved surfaces resembling broken glass. Such breaks are called *conchoidal fractures* (FIGURE 2.17). Still other minerals exhibit fractures that produce splinters or fibers that are referred to as *splintery* and *fibrous fracture*, respectively.

Density and Specific Gravity

Density, an important property of matter, is defined as mass per unit volume. Mineralogists often use a related measure called **specific gravity** to describe the density of minerals. Specific gravity is a number representing the ratio of a mineral's weight to the weight of an equal volume of water.

Most common rock-forming minerals have a specific gravity of between 2 and 3. For example, quartz has a specific gravity of 2.65. By contrast, some metallic minerals such as pyrite, native copper, and magnetite are more than twice as dense and thus have more than twice the specific gravity as quartz. Galena, an ore of lead, has a specific gravity of roughly 7.5, whereas the specific gravity of 24-karat gold is approximately 20.

With a little practice, you can estimate the specific gravity of a mineral by hefting it in your hand. Ask yourself, does this mineral feel about as "heavy" as similar sized rocks you have handled? If the answer is "yes," the specific gravity of the sample will likely be between 2.5 and 3.

Other Properties of Minerals

In addition to the properties discussed thus far, some minerals can be recognized by other distinctive properties. For example, halite is ordinary salt, so it can be quickly identified through taste. Talc and graphite both have distinctive feels; talc feels soapy, and graphite feels greasy. Further, the streaks of many sulfur-bearing minerals emit odors like rotten eggs. A few minerals, such as magnetite, have a high iron content and can be picked up with a magnet, while some varieties (lodestone) are natural magnets and will pick up small iron-based objects such as pins and paper clips (see Figure 2.28A, p. 57).

DID YOU KNOW?

One of the world's heaviest cut and polished gemstones is the 22,892.5-carat golden-yellow topaz. Currently housed in the Smithsonian Institution, this roughly 10-pound gem is about the size of an automobile headlight and could hardly be used as a piece of jewelry, except perhaps by an elephant.

Moreover, some minerals exhibit special optical properties. For example, when a transparent piece of calcite is placed over printed text, the letters appear twice. This optical property is known as *double refraction* (FIGURE 2.18).

One very simple chemical test involves placing a drop of dilute hydrochloric acid from a dropper bottle onto a freshly broken mineral surface. Using this technique, certain minerals, called carbonates, will effervesce (fizz) as carbon dioxide gas is released (FIGURE 2.19). This test is especially

FIGURE 2.18 Double refraction illustrated by the mineral calcite *(Photo by Chip Clark)*

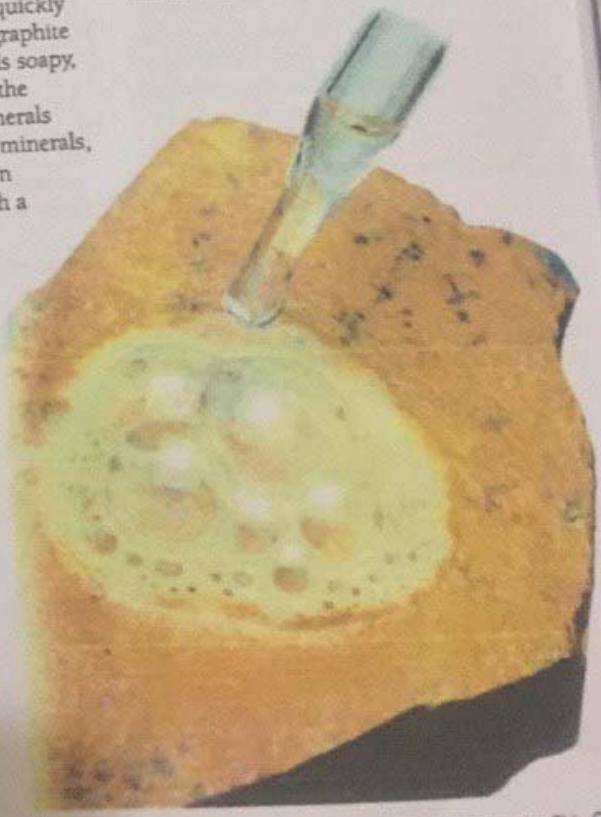


FIGURE 2.19 Calcite reacting with a weak acid *(Photo by Chip Clark)*

useful in identifying the common carbonate mineral calcite.

CONCEPT CHECK 2.5

1. Why is color not always a useful property in mineral identification? Give an example of a mineral that supports your answer.
2. What is meant when we refer to a mineral's tenacity? List four terms that describe tenacity.
3. Why does quartz lack cleavage?

Mineral Groups

Matter and Minerals
ESSENTIALS OF GEOLOGY **Minerals**

Over 4000 minerals have been named, and several new ones are identified each year. Fortunately, for students who are beginning to study minerals, no more than a few dozen are abundant! Collectively, these few make up most of the rocks of Earth's crust and, as such, are often referred to as the **rock-forming minerals.**

Although less abundant, many other minerals are used extensively in the manufacture of products and are called *economic minerals.* However, rock-forming minerals and economic minerals are not mutually exclusive groups. When found in large deposits, some rock-forming minerals are economically significant. One example is the mineral calcite, which is the primary component of the sedimentary rock limestone and has many uses including being used in the production of cement.

It is worth noting that *only eight elements* make up the vast majority of the rock-forming minerals and represent more than 98 percent (by weight) of the continental crust (FIGURE 2.20). These elements, in order of abundance, are oxygen (O), silicon (Si), aluminum (Al), iron (Fe), calcium (Ca), sodium (Na), potassium (K), and magnesium (Mg). As shown in Figure 2.20, silicon and oxygen are by far the most common elements in Earth's crust. Furthermore, these two elements readily combine to form the basic "building block" for the most common mineral group, the **silicates.** More than 800 silicate minerals are known, and they account for more than 90 percent of Earth's crust.

Because other mineral groups are far less abundant in Earth's crust than the silicates, they are often grouped together under the heading **nonsilicates.** Although not as common as silicates, some nonsilicate minerals are very important economically. They provide us with iron and aluminum to build our automobiles, gypsum for plaster and drywall for home construction, and copper wire that carries electricity and connects us to the Internet. Some common nonsilicate mineral groups include the carbonates, sulfates, and halides. In addition to their economic importance, these mineral groups include members that are major constituents in sediments and sedimentary rocks.

We first discuss the most common mineral group, the silicates, and then consider some of the prominent nonsilicate mineral groups.

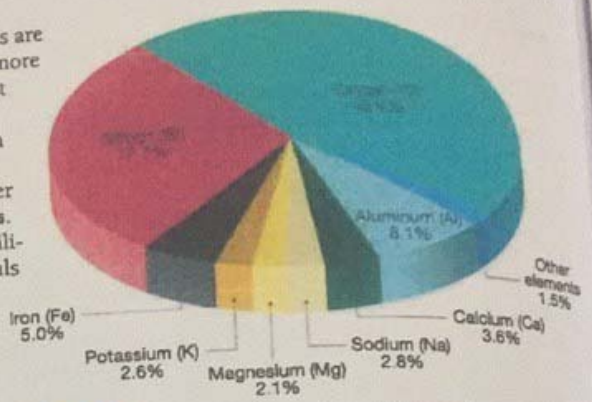


FIGURE 2.20 Relative abundance of the eight most abundant elements in the continental crust.

CONCEPT CHECK 2.6

1. List the eight most common elements in Earth's crust in order of abundance (most to least).
2. Explain the difference between the terms *silicon* and *silicate.*

The Silicates

Every silicate mineral contains the two most abundant elements of Earth's crust, oxygen and silicon. Further, most contain one or more of the other common elements. Together, these elements give rise to hundreds of silicate minerals with a wide variety of properties, including hard quartz, soft talc, sheet-like mica, fibrous asbestos, green olivine, and blood-red garnet.

Silicate Structures

All silicate minerals have the same fundamental building block, the **silicon–oxygen tetrahedron** (SiO_4^{4-}) This structure consists of four oxygen ions (each 2^-) that are covalently bonded to one comparatively small silicon ion (4^+) forming a *tetrahedron*—a pyramid shape with four identical faces (FIGURE 2.21). These tetrahedra are not chemical compounds, but rather complex ions (SiO_4^{4-}) having a net charge of -4. To become electrically balanced, these complex ions bond to other positively charged metal ions. Specifically, each O^{2-} has one of its valence electrons bonding with the Si^{4+}

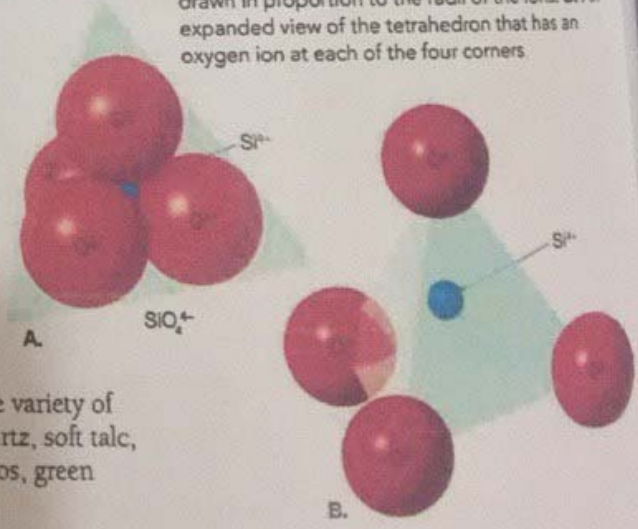


FIGURE 2.21 Two representations of the silicon–oxygen tetrahedron. **A.** The four large spheres represent oxygen ions, and the blue sphere represents a silicon ion. The spheres are drawn in proportion to the radii of the ions. **B.** An expanded view of the tetrahedron that has an oxygen ion at each of the four corners.

located at the center of the tetrahedron. The remaining 1^- charge on each oxygen is available to bond with another positive ion or with the silicon ion in an adjacent tetrahedron.

MINERALS WITH INDEPENDENT TETRAHEDRA. One of the simplest silicate structures consists of independent tetrahedra that have their four oxygen ions bonded to positive ions, such as Mg^{2+}, Fe^{2+}, and Ca^{2+}. The mineral olivine, with the formula $MgFe_2SiO_4$ is a good example. In olivine, magnesium (Mg^{2+}) and/or iron (Fe^{2+}) ions pack between comparatively large independent SiO_4 tetrahedra, forming a dense three-dimensional structure. Garnet, another common silicate, is also composed of independent tetrahedra ionically bonded by positive ions. Both olivine and garnet form dense, hard, equidimensional crystals that lack cleavage.

MINERALS WITH CHAIN OR SHEET STRUCTURES. One reason for the great variety of silicate minerals is the ability of SiO_4 tetrahedra to link to one another in a variety of configurations. This important phenomenon, called **polymerization**, is achieved by the sharing of one, two, three, or all four of the oxygen atoms with adjacent tetrahedra. Vast numbers of tetrahedra join together to form single chains, double chains, sheet structures, or three-dimensional frameworks as shown in FIGURE 2.22.

To see how oxygen atoms are shared between adjacent tetrahedra, select one of the silicon ions (small blue spheres) near the middle of the single-chain shown in Figure 2.22B. Notice that this silicon ion is completely surrounded by four larger oxygen ions. Also notice that, of the four oxygen atoms, half are bonded to two silicon atoms, whereas the other two are not shared in this manner. It is the linkage across the shared oxygen ions that join the tetrahedra into a chain structure. Now examine a silicon ion near the middle of the sheet structure (Figure 2.22D) and count the number of shared and unshared oxygen ions surrounding it. As you likely observed, the sheet structure is the result of three of the four oxygen atoms being shared by adjacent tetrahedra.

MINERALS WITH THREE-DIMENSIONAL FRAMEWORKS. In the most common silicate structure, all four oxygen ions are shared, producing a complex three-dimensional framework (Figure 2.22E). Quartz, a hard, durable mineral, has the simplest structure in which all of the oxygens are shared. Because its structure is electrically neutral, quartz (SiO_2) contains no positive ions—other than silicon.

The ratio of oxygen ions to silicon ions differs in each type of silicate structure. In independent tetrahedra (SiO_4) there are four oxygen ions for every silicon ion. In single chains, the oxygen-to-silicon ratio is 3:1 (SiO_3), and in three-dimensional frameworks as found in quartz the ratio is 2:1 (SiO_2). As more oxygen ions are shared, the percentage of silicon in the structure increases. Silicate minerals are, therefore, described as having a low or high silicon content based on their ratio of oxygen to silicon. Minerals with three-dimensional structures in which all four oxygen ions are shared have the highest silicon content. Minerals composed of independent tetrahedra have the lowest. This difference in silicon content is important, as you will see in Chapter 3.

Joining Silicate Structures

Except for quartz (SiO_2) the basic structure (chains, sheets, or three-dimensional frameworks) of most silicate minerals has a net negative charge. Therefore, metal ions are required to bring the overall charge into balance and to serve as the "mortar" that holds these structures together. The positive ions that most often link silicate structures

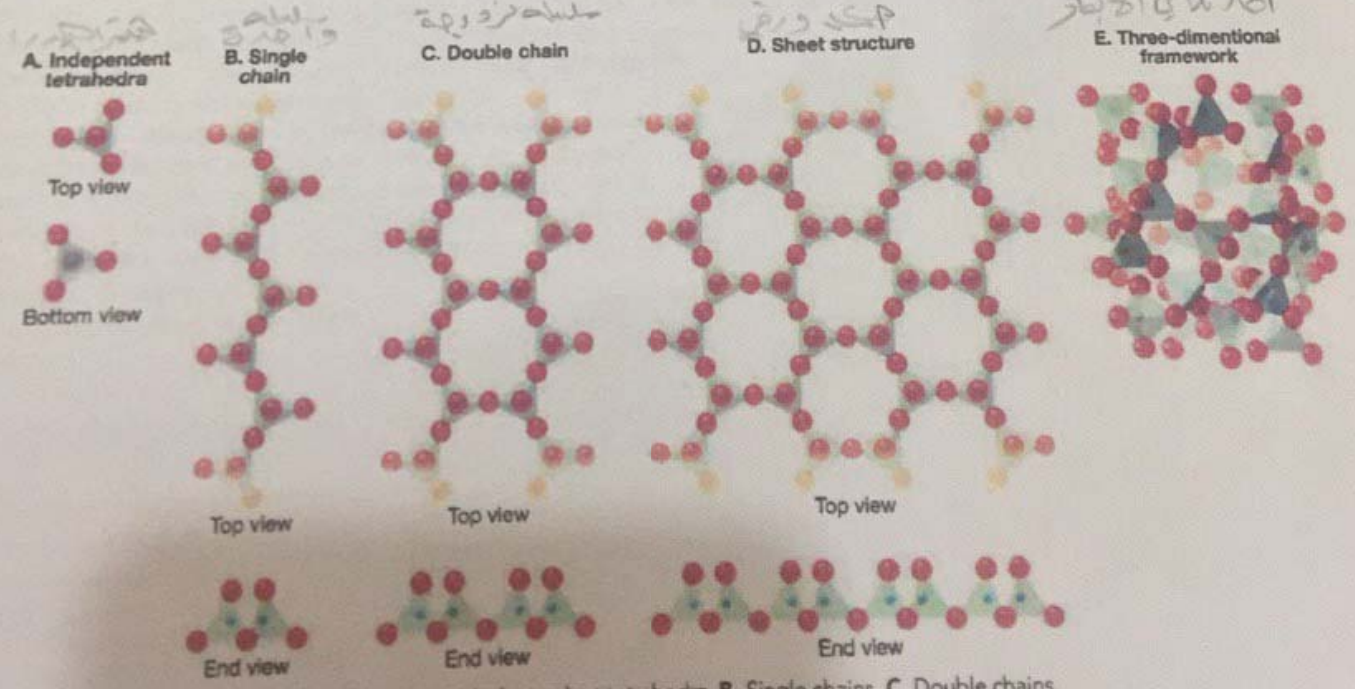


FIGURE 2.22 Five types of silicate structures. A. Independent tetrahedra. B. Single chains. C. Double chains. D. Sheet structures. E. Three-dimensional framework.

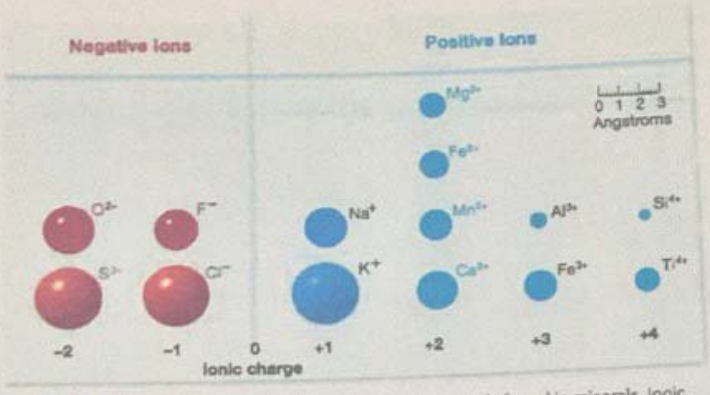


FIGURE 2.23 The relative sizes and charges of ions commonly found in minerals. Ionic radii are usually expressed in angstroms (1 angstrom equals 10^{-8} cm)

are iron (Fe^{2+}), magnesium (Mg^{2+}), potassium (K^{1+}), sodium (Na^{1+}), aluminum (Al^{3+}), and calcium (Ca^{2+}). These positively charged ions bond with the unshared oxygen ions that occupy the corners of the silicate tetrahedra (**FIGURE 2.23**).

As a general rule, the hybrid covalent bonds between silicon and oxygen are stronger than the ionic bonds that hold one silicate structure to the next. Consequently, properties such as cleavage, and to some extent hardness, are controlled by the nature of the silicate framework. Quartz (SiO_2), which has only silicon–oxygen bonds, has great hardness and lacks cleavage, mainly because of equally strong bonds in all directions. By contrast, the mineral talc (the source of talcum powder), has a sheet structure. Magnesium ions occur between the sheets and weakly join them together. The slippery feel of talcum powder is due to the silicate sheets sliding relative to one another, in much the same way sheets of carbon atoms slide in graphite, giving it its lubricating properties.

Recall that atoms of similar size can substitute freely for one another without altering a mineral's structure. For example, in the mineral olivine, iron (Fe^{2+}) and magnesium (Mg^{2+}) substitute for each other. This also holds true for the third most common element in Earth's crust, aluminum (Al^{3+}), which often substitutes for silicon (Si) in the center of silicon–oxygen tetrahedra.

Because most silicate structures will readily accommodate two or more different positive ions at a given bonding site, individual specimens of a particular mineral may contain varying amounts of certain elements. As a result, many silicate minerals form a *mineral group* that exhibits a range of compositions between two end members. Examples include the olivines, pyroxenes, amphiboles, micas, and feldspars.

CONCEPT CHECK 2.7

1. Sketch a silicon–oxygen tetrahedron.
2. Explain the following statement:
   Silicate minerals with three-dimensional structures have the highest silicon content, while those composed of independent tetrahedra have the lowest.

Common Silicate Minerals

Matter and Minerals
ESSENTIALS OF GEOLOGY **Mineral Groups**

The major groups of silicate minerals and common examples are given in **FIGURE 2.24**. The feldspars are, by far, the most plentiful silicate group, comprising more than 50 percent of Earth's crust. Quartz, the second most abundant mineral in the continental crust, is the only common mineral made completely of silicon and oxygen.

Most silicate minerals form when molten rock cools and crystallizes. Cooling can occur at or near Earth's surface (low temperature and pressure) or at great depths (high temperature and pressure). The environment during crystallization and the chemical composition of the molten rock determine, to a large degree, which

Uncut emerald crystal, a variety of the mineral beryl. (Photo by Jeffrey Scovil)

Mineral/Formula	Cleavage	Silicate Structure	Example
Olivine group $(Mg, Fe)_2SiO_4$	None	Single tetrahedron	Olivine
Pyroxene group (Augite) $(Mg,Fe)SiO_3$	Two planes at 90°	Single chains	Augite
Amphibole group (Hornblende) $Ca_2(Fe,Mg)_5Si_8O_{22}(OH)_2$	Two planes at 60° and 120°	Double chains	Hornblende
Micas Biotite $K(Mg,Fe)_3AlSi_3O_{10}(OH)_2$	One plane	Sheets	Biotite
Micas Muscovite $KAl_2(AlSi_3O_{10})(OH)_2$	One plane	Sheets	Muscovite
Feldspars Potassium feldspar (Orthoclase) $KAlSi_3O_8$	Two planes at 90°	Three-dimensional networks	Potassium feldspar
Feldspars Plagioclase $(Ca,Na)AlSi_3O_8$	Two planes at 90°	Three-dimensional networks	
Quartz SiO_2	None	Three-dimensional networks	Quartz

FIGURE 2.24 Common silicate minerals. Note that the complexity of the silicate structure increases from top to bottom. (Photos by Dennis Tasa and E. J. Tarbuck)

DID YOU KNOW?
Most "scoopable" kitty litters sold on the market today contain a naturally occurring material called *bentonite*. Bentonite is largely composed of highly absorbent clay minerals that swell and clump up in the presence of moisture. This allows kitty waste to be isolated and scooped out, leaving behind clean litter.

minerals are produced. For example, the silicate mineral olivine crystallizes at high temperatures, whereas quartz crystallizes at much lower temperatures.

In addition, some silicate minerals form at Earth's surface from the weathered products of other silicate minerals. Still others are formed under the extreme pressures associated with mountain building. Each silicate mineral, therefore, has a structure and a chemical composition that *indicate the conditions under which it formed*. By carefully examining the mineral constituents of rocks, geologists can usually determine the circumstances under which the rocks formed.

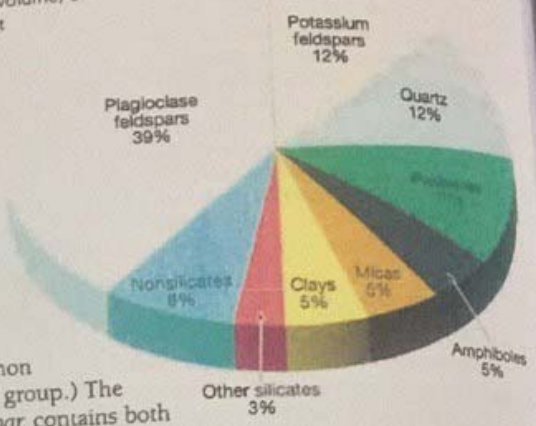
We will now examine some of the most common silicate minerals, which we divide into two major groups on the basis of their chemical makeup.

The Light Silicates

The **light (or nonferromagnesian) silicates** are generally light in color and have a specific gravity of about 2.7, which is considerably less than the dark (ferromagnesian) silicates. These differences are mainly attributable to the presence or absence of iron and magnesium. The light silicates contain varying amounts of aluminum, potassium, calcium, and sodium rather than iron and magnesium.

FELDSPAR GROUP. *Feldspar*, the most common mineral group, can form under a wide range of temperatures and pressures, which partially accounts for its abundance (FIGURE 2.25). All feldspars have similar physical properties. They have two planes of cleavage meeting at or near 90-degree angles, are relatively hard (6 on the Mohs scale), and have a luster that ranges from glassy to pearly. As one component in a rock, feldspar crystals can be identified by their rectangular shape and rather smooth shiny faces (see Figure 2.24).

FIGURE 2.25 Estimated percentages (by volume) of the most common minerals in Earth's crust



Two different feldspar structures exist. One group of feldspar minerals contains potassium ions in its structure and is therefore referred to as *potassium feldspar*. (*Orthoclase* and *microcline* are common members of the potassium feldspar group.) The other group, called *plagioclase feldspar*, contains both sodium and calcium ions that freely substitute for one another depending on the environment during crystalization.

Potassium feldspar is usually light cream, salmon pink, or occasionally bluish-green in color. The plagioclase feldspars, on the other hand, range in color from white to medium gray. However, color should not be used to distinguish these groups. The only way to distinguish the feldspars physically is to look for a multitude of fine parallel lines, called *striations*. Striations are found on some cleavage planes of plagioclase feldspar but are not present on potassium feldspar (FIGURE 2.26).

DID YOU KNOW?
Mica crystals have been known to grow to great size. One found in Russia's Ural Mountains was 54 square feet in area and more than 1.5 feet thick.

QUARTZ. *Quartz* is the only common silicate mineral consisting entirely of silicon and oxygen. As such, the term *silica* is applied to quartz, which has the chemical formula SiO_2. Because the structure of quartz contains a ratio of two oxygen ions (O^{2-}) for every silicon ion (Si^{4+}), no other positive ions are needed to attain neutrality.

In quartz, a three-dimensional framework is developed through the complete sharing of oxygen by adjacent silicon atoms. Thus, all of the bonds in quartz are of the strong silicon–oxygen type. Consequently, quartz is hard, resistant to weathering, and does not have cleavage. When broken, quartz generally exhibits conchoidal fracture (see Figure 2.17). When pure, quartz is clear and, if allowed to grow without interference, will develop hexagonal crystals that develop pyramid-shaped ends. However, like most other clear minerals, quartz is often colored by inclusions of various ions (impurities) and forms without developing good crystal faces. The most common varieties of quartz are milky (white), smoky (gray), rose (pink), amethyst (purple), and rock crystal (clear) (see Figure 2.10).

MUSCOVITE. *Muscovite* is a common member of the mica family. It is light in

FIGURE 2.26 These parallel lines, called striations, are a distinguishing characteristic of plagioclase feldspar. (Photo by E. J. Tarbuck)

color and has a pearly luster (see Figure 2.15). Like other micas, muscovite has excellent cleavage in one direction. In thin sheets, muscovite is clear, a property that accounts for its use as window "glass" during the Middle Ages. Because muscovite is very shiny, it can often be identified by the sparkle it gives a rock. If you have ever looked closely at beach sand, you may have seen the glimmering brilliance of the mica flakes scattered among the other sand grains.

Basalt is a common igneous rock composed of a large percentage of dark silicate minerals. *(Photo by Roger Rassmeyer/CORBIS)*

DID YOU KNOW?

Although wood pulp is the main ingredient for making newsprint, many higher grades of paper contain large quantities of clay minerals. In fact, each page of this textbook consists of about 25 percent clay (the mineral kaolinite). If all the clay in this book rolled into a ball, it would be slightly smaller than a tennis ball.

CLAY MINERALS. *Clay* is a term used to describe a category of complex minerals that, like the micas, have a sheet structure. Unlike other common silicates, such as quartz and feldspar, most clay minerals originate as products of the chemical weathering of other silicate minerals. Thus, clay minerals make up a large percentage of the surface material we call soil. Because of the importance of soil in agriculture, and because of its role as a supporting material for buildings, clay minerals are extremely important to humans. In addition, clays account for nearly half the volume of sedimentary rocks. Clay minerals are generally very fine grained, which makes identification difficult, unless studied microscopically. Their layered structure and weak bonding between layers give them a characteristic feel when wet. Clays are common in shales, mudstones, and other sedimentary rocks.

One of the most common clay minerals is *kaolinite*, which is used in the manufacture of fine china and as a coating for high-gloss paper, such as that used in this textbook. Further, some clay minerals absorb large amounts of water, which allows them to swell to several times their normal size. These clays have been used commercially in a variety of ingenious ways, including as an additive to thicken milkshakes in fast-food restaurants.

The Dark Silicates

The **dark (or ferromagnesian) silicates** are those minerals containing ions of iron (iron = *ferro*) and/or magnesium in their structure. Because of their iron content, ferromagnesian silicates are dark in color and have a greater specific gravity, between

3.2 and 3.6, than nonferromagnesian silicates. The most common dark silicate minerals are olivine, the pyroxenes, the amphiboles, dark mica (biotite), and garnet.

OLIVINE GROUP. *Olivine* is a family of high-temperature silicate minerals that are black to olive green in color and have a glassy luster and a conchoidal fracture (see Figure 2.24). Transparent olivine is occasionally used as a gemstone called peridot. Rather than developing large crystals, olivine commonly forms small, rounded crystals that give olivine-rich rocks a granular appearance. Olivine and related forms are thought to constitute up to 50 percent of Earth's upper mantle.

PYROXENE GROUP. The *pyroxenes* are a group of complex minerals that are important components in dark colored igneous rocks. The most common member, *augite*, is a black, opaque mineral with two directions of cleavage that meet at nearly a 90-degree angle. Augite is one of the dominant minerals in basalt, a common igneous rock of the oceanic crust and volcanic areas on the continents.

AMPHIBOLE GROUP. *Hornblende* is the most common member of a chemically complex group of minerals called *amphiboles* (see Figure 2.24). Hornblende is usually dark green to black in color, and except for its cleavage angles, which are about 60 degrees and 120 degrees, it is very similar in appearance to augite. In a rock, hornblende often forms elongated crystals. This helps distinguish it from pyroxene, which forms rather blocky crystals. Hornblende is found in igneous rocks, where it often makes up the dark portion of an otherwise light-colored rock.

BIOTITE. *Biotite* is the dark, iron-rich member of the mica family (see Figure 2.24). Like other micas, biotite possesses a sheet structure that gives it excellent cleavage in one direction. Biotite also has a shiny black appearance that helps distinguish it from the other dark ferromagnesian minerals. Like hornblende, biotite is a common constituent of igneous rocks, including the rock granite.

GARNET. *Garnet* is similar to olivine in that its structure is composed of individual tetrahedra linked by metallic ions. Also like olivine, garnet has a glassy luster, lacks cleavage, and exhibits conchoidal fracture. Although the colors of garnet are varied, this mineral is most often brown to deep red. Garnet readily forms equidimensional crystals that are most commonly found in metamorphic rocks. When garnets are transparent, they may be used as gemstones.

CONCEPT CHECK 2.8

1. What do ferromagnesian minerals have in common? List three examples of ferromagnesian minerals.
2. What do muscovite and biotite have in common? How do they differ?
3. Should color be used to distinguish between orthoclase and plagioclase feldspar? What is the most effective means of distinguishing them?

Important Nonsilicate Minerals

Matter and Minerals

ESSENTIALS OF GEOLOGY **Mineral Groups**

Nonsilicate minerals are typically divided into groups, based on the negatively charged ion or complex ion that the members have in common (TABLE 2.1). For example, the *oxides* contain the negative oxygen ion (O^{2-}), which is bonded to one or more kinds of positive ions. Thus, within each mineral group, the basic structure and type of bonding is similar. As a result, the minerals in each group have similar physical properties that are useful in mineral identification.

Although the nonsilicates make up only about 8 percent of Earth's crust, some minerals, such as gypsum, calcite, and halite, occur as constituents in sedimentary rocks in significant amounts. Furthermore, many others are important economically. Table 2.1 lists some of the nonsilicate mineral groups and a few examples of each. A brief discussion of a few of the more common nonsilicate minerals follows.

DID YOU KNOW?

The material we commonly call mother-of-pearl is a carbonate mineral called aragonite. Mother-of-pearl is produced by a large group of marine organisms commonly called clams (mollusks).

Some of the most common nonsilicate minerals belong to one of three classes of minerals—the carbonates (CO_3^{2-}), the sulfates (SO_4^{2-}), and the halides (Cl^{1-}, F^{1-}, Br^{1-}). The carbonate minerals are much simpler structurally than the silicates. This mineral group is composed of the carbonate ion (CO_3^{2-}) and one or more kinds of positive ions. The two most common carbonate minerals are *calcite*, $CaCO_3$ (calcium carbonate), and *dolomite*, $CaMg(CO_3)_2$ (calcium/magnesium carbonate). Because these minerals are similar both physically and chemically, they are difficult to distinguish from each other. Both have a vitreous luster, a hardness between 3 and 4, and nearly perfect rhombic cleavage. They can, however, be distinguished by using dilute hydrochloric acid. Calcite reacts vigorously with this acid, whereas dolomite reacts much more slowly. Calcite and dolomite are usually found together as the primary constituents in the sedimentary rocks limestone and dolostone. When calcite is the dominant mineral, the rock is called *limestone*, whereas *dolostone* results from a predominance of dolomite. Limestone has many uses, including as road aggregate, as building stone, and as the main ingredient in Portland cement.

Two other nonsilicate minerals frequently found in sedimentary rocks are *halite* and *gypsum*. Both minerals are commonly found in thick layers that are the last vestiges of ancient seas that have

DID YOU KNOW?

The mineral pyrite is commonly known as "fool's gold," because its golden-yellow color can lead to its being mistaken for gold. The name pyrite is derived from the Greek *pyros* ("fire"), because it gives off sparks when struck sharply.

long since evaporated (FIGURE 2.27). Like limestone, both are important nonmetallic resources. Halite is the mineral name for common table salt (NaCl). Gypsum ($CaSO_4 \cdot 2H_2O$), which is calcium sulfate with water bound into the structure, is the mineral of which plaster and other similar building materials are composed.

Most nonsilicate mineral classes contain members that are prized for their economic value. This includes the oxides, whose members hematite and magnetite are important ores of iron (FIGURE 2.28). Also significant are the sulfides, which are basically compounds of sulfur (S) and one or more metals. Examples of important sulfide minerals include galena (lead), sphalerite (zinc), and chalcopyrite (copper). In addition, native elements, including gold, silver, and carbon (diamonds), plus a host of other nonsilicate minerals—fluorite (flux in making steel), corundum (gemstone, abrasive), and uraninite (a uranium source)—are important economically.

CONCEPT CHECK 2.9

1. List six common nonsilicate mineral groups. What key ion(s) or element(s) define each group?
2. What simple test can be used to distinguish calcite from dolomite?
3. List eight common nonsilicate minerals and their economic uses.

TABLE 2.1
Common Nonsilicate Mineral Groups

Mineral Groups [key ion(s) or element(s)]	Mineral Name	Chemical Formula	Economic Use
Carbonates (CO_3^{2-})	Calcite	$CaCO_3$	Portland cement, lime
	Dolomite	$CaMg(CO_3)_2$	Portland cement, lime
Halides (Cl^{1-}, F^{1-}, Br^{1-})	Halite	NaCl	Common salt
	Fluorite	CaF_2	Used in steelmaking
	Sylvite	KCl	Fertilizer
Oxides (O^{2-})	Hematite	Fe_2O_3	Ore of iron, pigment
	Magnetite	Fe_3O_4	Ore of iron
	Corundum	Al_2O_3	Gemstone, abrasive
	Ice	H_2O	Solid form of water
Sulfides (S^{2-})	Galena	PbS	Ore of lead
	Sphalerite	ZnS	Ore of zinc
	Pyrite	FeS_2	Sulfuric acid production
	Chalcopyrite	$CuFeS_2$	Ore of copper
	Cinnabar	HgS	Ore of mercury
Sulfates (SO_4^{2-})	Gypsum	$CaSO_4 \cdot 2H_2O$	Plaster
	Anhydrite	$CaSO_4$	Plaster
	Barite	$BaSO_4$	Drilling mud
Native elements (single elements)	Gold	Au	Trade, jewelry
	Copper	Cu	Electrical conductor
	Diamond	C	Gemstone, abrasive
	Sulfur	S	Sulfa drugs, chemicals
	Graphite	C	Pencil lead, dry lubricant
	Silver	Ag	Jewelry, photography
	Platinum	Pt	Catalyst

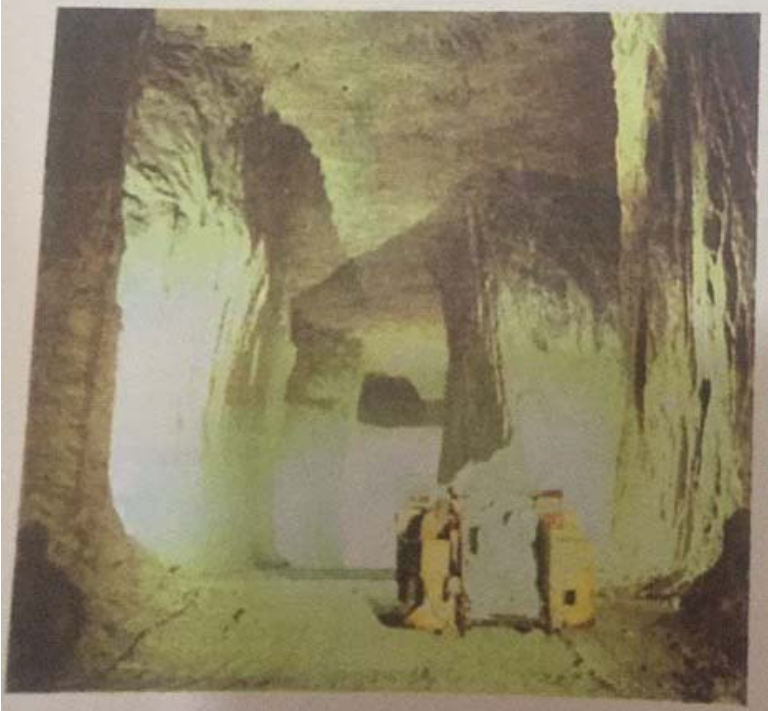


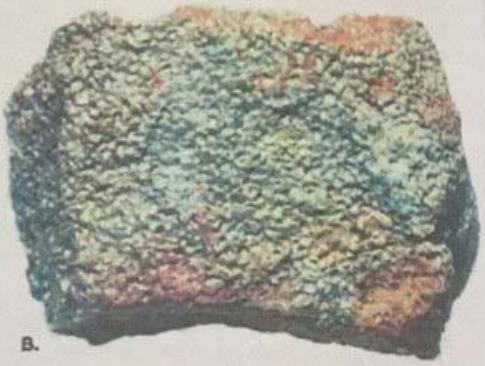
FIGURE 2.27 Thick bed of halite (salt) at an underground mine in Grand Saline, Texas. Note person for scale. *(Photo by Tom Bochsler)*

Mineral Resources

Earth's crust and oceans are the source of a wide variety of useful and essential substances. In fact, practically every manufactured product contains materials obtained from minerals. Table 2.1 lists important mineral groups in this category.

Mineral resources are the endowment of useful minerals ultimately available commercially. Resources include already identified deposits from which minerals can be extracted profitably, called **reserves**, as well as known deposits that are not yet economically or technologically recoverable. Deposits inferred to exist, but not yet discovered, are also considered mineral resources.

A.



B.

FIGURE 2.28 Magnetite **A.** and hematite **B.** are both oxides and are both important ores of iron (Photos by E. J. Tarbuck)

The term **ore** is used to denote those useful metallic minerals that can be mined at a profit (Figure 2.28). In common usage, the term ore is also applied to some nonmetallic minerals such as fluorite and sulfur. However, materials used for such purposes as building stone, road aggregate, abrasives, ceramics, and fertilizers are not usually called ores; rather, they are classified as industrial rocks and minerals.

Recall that more than 98 percent of Earth's crust is composed of only eight elements, and except for oxygen and silicon, all other elements make up a relatively small fraction of common crustal rocks (see Figure 2.20). Indeed, the natural concentrations of many elements are exceedingly small. A deposit containing the average percentage of a valuable element such as gold has no economic value, because the cost of extracting it greatly exceeds the value of the gold that could be recovered.

To have economic value, an element must be concentrated above the level of its average crustal abundance. For example, copper makes up about 0.0135 percent of the crust. For a deposit to be considered as copper ore, it must contain a concentration that is about 100 times this amount. Aluminum, on the other hand, represents 8.13 percent of the crust and can be extracted profitably when it is found in concentrations only about four times its average crustal percentage.

It is important to realize that a deposit may become profitable to extract or lose its profitability because of economic changes. If demand for a metal increases and prices rise sufficiently, the status of a previously unprofitable deposit changes, and it becomes an ore. The status of unprofitable deposits may also change if a technological advance allows the ore to be extracted at a lower cost than before.

Conversely, changing economic factors can turn a once profitable ore deposit into an unprofitable deposit that can no longer be called an ore. This situation was illustrated at the copper mining operation located at Bingham Canyon, Utah, one of the largest open-pit mines on Earth (**FIGURE 2.29**). Mining was halted there in 1985 because outmoded equipment had driven the cost of extracting the copper beyond the current selling price. The owners responded by replacing an antiquated 1000-car railroad with conveyor belts and pipelines for transporting the ore and waste. These devices achieved a cost reduction of nearly 30 percent and returned this mining operation to profitability.

Over the years, geologists have been keenly interested in learning how natural processes produce localized concentrations of essential minerals. One well-established fact is that occurrences of valuable mineral resources are closely related to the rock cycle. That is, the mechanisms that generate igneous, sedimentary, and metamorphic rocks, including the processes of weathering and erosion, play a major role in producing concentrated accumulations of useful elements.

Moreover, with the development of the theory of plate tectonics, geologists have added another tool for understanding the processes by which one rock is transformed into another. As these rock-forming processes are examined in the following chapters, we will consider their role in producing some of our important mineral resources.

DID YOU KNOW?

Gypsum, a white-to-transparent mineral, was first used as a building material in Anatolia (present-day Turkey) around 6000 BC. It is also found on the interiors of the great pyramids in Egypt, which were erected in about 3700 BC. Today, an average new American home contains more than 7 metric tons of gypsum in the form of 6000 square feet of wallboard.

CONCEPT CHECK 2.10

1. Contrast a mineral *resource* and a mineral *reserve*.
2. What might cause a mineral deposit that had not been considered an ore to become reclassified as an ore?

CHAPTER TWO

Matter and Minerals in Review

A *mineral* is any naturally occurring inorganic solid that possesses an orderly crystalline structure and can be represented by a chemical formula. Most *rocks* are aggregates composed of two or more minerals.

All matter, including minerals, is composed of minute, indivisible particles called *atoms*—the building blocks of minerals. Each atom has a *nucleus*, which contains *protons* (particles with positive electrical charges) and *neutrons* (particles with neutral electrical charges). Surrounding the nucleus of an atom in regions called *principal shells*, are *electrons*, which have negative electrical charges. The number of protons in an atom's nucleus determines its *atomic number* and the name of the element. An *element* is a large collection of electrically neutral atoms, all having the same atomic number.

Atoms combine with each other to form more complex substances called *compounds*. Atoms bond by gaining, losing, or sharing electrons with other atoms. In *ionic bonding*, one or more electrons are transferred from one atom to another, giving the atoms a net positive or negative charge. The resulting electrically charged atoms are called *ions*. Ionic compounds consist of oppositely charged ions assembled in a regular, crystalline structure that allows for the maximum attraction of ions, given their sizes. Another type of bond, the *covalent bond*, is produced when atoms share electrons.

Isotopes are variants of the same element that have a different *mass number* (the total number of neutrons plus protons found in an atom's nucleus). Some isotopes are unstable and disintegrate naturally through a process called *radioactivity*.

The properties of minerals include *crystal shape (habit)*, *luster*, *color*, *streak*, *tenacity*, *hardness*, *cleavage*, *fracture*, and *density* or *specific gravity*. In addition, a number of special physical and chemical properties (*taste*, *smell*, *elasticity*, *feel*, *magnetism*, *double refraction*, and *chemical reaction to hydrochloric acid*) are useful in identifying certain minerals. Each mineral has a unique set of properties that can be used for identification.

Of the nearly 4000 minerals, no more than a few dozen make up most of the rocks of Earth's crust and, as such, are classified as rock-forming minerals. Eight elements (oxygen, silicon, aluminum, iron, calcium, sodium, potassium, and magnesium) make up the bulk of these minerals and represent over 98 percent (by weight) of Earth's continental crust.

The most common mineral group is the *silicates*. All silicate minerals have the negatively charged *silicon–oxygen tetrahedron* as their fundamental building block. In some silicate minerals the tetrahedra are joined in chains (the pyroxene and amphibole groups); in others, the tetrahedra are arranged into sheets (the micas—biotite and muscovite) or three-dimensional networks (the feldspars and quartz). The tetrahedra and various silicate structures are often bonded together by the positive ions of iron, magnesium, potassium, sodium, aluminum, and calcium. Each silicate mineral has a structure and a chemical composition that indicates the conditions under which it formed.

The *nonsilicate* mineral groups, which contain several economically important minerals, include the *oxides* (e.g., the mineral hematite, mined for iron), *sulfides* (e.g., the mineral sphalerite, mined for zinc, and the mineral galena, mined for lead), *sulfates*, *halides*, and *native elements* (e.g., gold and silver). The more common nonsilicate rock-forming minerals include the *carbonate minerals*, calcite and dolomite. Two other nonsilicate minerals frequently found in sedimentary rocks are halite and gypsum.

Mineral resources are the endowment of useful minerals ultimately available commercially. Resources include already identified deposits from which minerals can be extracted profitably, called *reserves*, as well as known deposits that are not yet economically or technologically recoverable. Deposits inferred to exist, but not yet discovered, are also considered mineral resources. The term *ore* is used to denote those useful metallic minerals that can be mined for a profit, as well as some nonmetallic minerals, such as fluorite and sulfur, that contain useful substances.

MINERALS

DEFINITION

A mineral is a naturally occurring, crystalline, inorganic, homogeneous solid with a chemical composition that is either fixed or varies within certain fixed limits, and a characteristic internal structure manifested in its exterior form and physical properties.

MINERAL IDENTIFICATION

Common minerals are identified or recognized by testing them for general or specific physical properties. For example, the common substance table salt is actually a mineral composed of sodium chloride (NaCl) and bears the mineral name halite. The taste of halite is distinctive and is sufficient for identifying and distinguishing it from other substances such as sugar (not a mineral) that have a similar appearance. Chemical composition alone is not sufficient to identify minerals. The mineral graphite and the mineral diamond are both composed of a single element, carbon (C), but their physical properties are very different.

The taste test applied to halite is restrictive because it is the only mineral that can be so identified. Other minerals may have a specific taste, but theirs are different from that of halite. Other common minerals can be tested by visual inspection for the physical properties of *crystal form, cleavage,* or *color* or by using simple tools such as a knife blade or glass plate to test for the physical property of hardness.

The first step in learning how to identify common minerals is to become acquainted with the various physical properties that individually or collectively characterize a mineral specimen.

PROPERTIES OF MINERALS

The physical properties of minerals are those that can be observed generally in all minerals. They include such common features as luster, color, hardness, cleavage, streak, and specific gravity. *Special properties* are those that are found in only a few minerals. These include magnetism, double refraction, taste, odor, feel, and chemical reaction with acid.

In your work in the laboratory, use the hand specimens sparingly when applying tests for the various properties.

General Physical Properties

LUSTER

The appearance of a fresh mineral surface in reflected light is its luster. A mineral that looks like a metal is said to have a *metallic luster.* Minerals that are *nonmetallic* are described by one of the following adjectives: *vitreous* (having the luster of glass); *resinous* (having the luster of resin); *pearly; silky; dull* or *earthy* (not bright or shiny).

Your laboratory instructor will display examples of minerals that possess these various lusters.

COLOR

The color of a mineral is determined by examining a fresh surface in reflected light. Color and luster are not the same. Some minerals are clear and transparent, thus colorless. The color or lack of color may be diagnostic in some minerals, but in others the color varies due either to a slight difference in chemical composition or to small amounts of impurities within the mineral (fig. 1.1).

Figure 1.1

The specimens in this photograph are all quartz. The difference in colors is due to various impurities. Clockwise from left: smoky quartz, quartz crystal, rose quartz, citrine quartz, amethyst quartz

HARDNESS

The hardness of a mineral is its resistance to abrasion. Hardness can be determined either by trying to scratch a mineral of unknown hardness with a substance of known hardness, or by using the unknown mineral to scratch a substance of known hardness. Hardness is determined on a relative scale called the *Mohs scale of hardness,* which consists of ten common minerals arranged in order of their increasing hardness (table 1.1). In the laboratory, convenient materials other than these ten specific minerals may be used for hardness determination.

In this manual, a mineral that scratches glass will be considered "hard," and one that does not scratch glass will be considered "soft." In making hardness tests on a glass plate, do not hold the glass in your hands; keep it firmly on the tabletop. If you think that you have made a scratch on the glass, try to rub the scratch off. What appears to be a scratch may only be some of the mineral that has rubbed off on the glass.

CLEAVAGE

Cleavage is the tendency of a mineral to break along definite planes of weakness that exist in the internal (atomic) structure of the mineral. Cleavage planes are related to the crystal system of the mineral and are always parallel to crystal faces or possible crystal faces. Cleavage may be conspicuous and is a characteristic physical property that is useful in mineral identification. It is almost impossible to break some minerals in such a way that cleavage planes do not develop. An example is calcite, with its rhombohedral cleavage.

Perfect cleavage describes cleavage planes that are very smooth and flat and that reflect light much like a mirror. Other descriptors such as *good, fair,* and *poor* are used to describe cleavage surfaces that are less well defined. Some minerals exhibit excellent crystal faces but have no cleavage; quartz is such a mineral.

The cleavage planes of some minerals such as calcite, muscovite (fig. 1.2), halite, and fluorite are so well developed that they are easily detected. In others, the cleavage surfaces may be so discontinuous as to escape detection by casual inspection. Before deciding that a mineral has no cleavage, turn it around in a strong light and observe whether there is some position in which the surface of the specimen "lights up," that is, reflects the light as if it were the reflecting surface of a dull mirror. If so, the mineral has cleavage, but the cleavage surface consists of several discontinuous parallel planes minutely separated, and rather than perfect cleavage it has good, fair, or poor cleavage. As will be noted in the discussion of crystal form, it is important to differentiate between cleavage planes and crystal faces (the actual breaking of a mineral crystal may be useful in making this differentiation).

In assigning the number of cleavage planes to a specimen, do not make the mistake of calling two parallel planes bounding the opposite sides of a specimen two cleavage planes. In this case, the specimen has two cleavage surfaces but only one plane of cleavage (i.e., one direction of cleavage). For example, halite has cubic cleavage, thus six sides, but only three planes of cleavage, because the six sides are made up of three parallel pairs of cleavage surfaces.

The angle at which two cleavage planes intersect is diagnostic. This angle can be determined by inspection. In most cases you will need to know whether the angle is 90 degrees, almost 90 degrees, or more or less than 90 degrees. The cleavage relationships that you will encounter during the course of your study of common minerals are tabulated for convenience in figure 1.3.

Some minerals exhibit the characteristic of *parting,* sometimes called false cleavage. Parting occurs along planes of weakness in the mineral, but usually the planes are more widely separated and often are due to twinning deformation or inclusions. Parting is not present in all specimens of a given mineral.

TABLE 1.1

Mineral Hardness According to the Mohs Scale (A) and Some Common Materials (B)

Hardness	A	B
1	Talc	
2	Gypsum	
2.5		Fingernail
3	Calcite	
3.5		Copper penny
4	Fluorite	
5	Apatite	
5–5.5		Knife blade
5.5		Glass plate
6	Feldspar	
6.5		Steel file
7	Quartz	
8	Topaz	
9	Corundum	
10	Diamond	

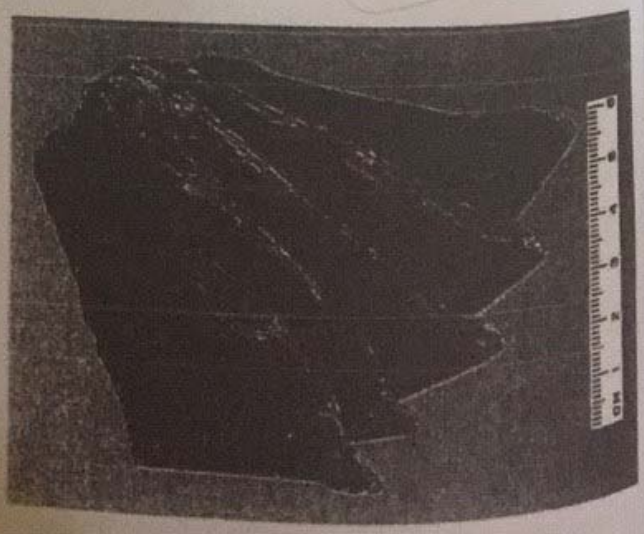


Figure 1.2

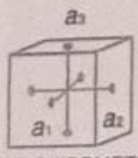
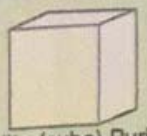
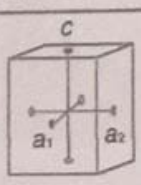
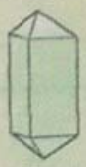
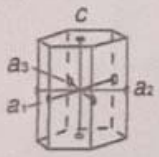
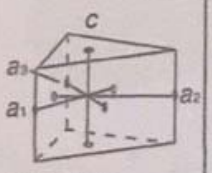
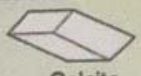
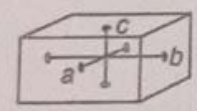
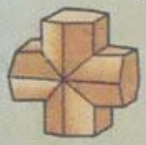
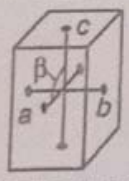
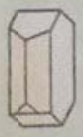
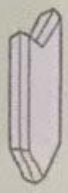
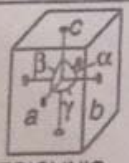
CRYSTAL SYSTEM	CHARACTERISTICS	EXAMPLES*
 CUBIC (ISOMETRIC)	Three mutually perpendicular axes, all of the same length ($a_1 = a_2 = a_3$). Fourfold axis of symmetry around a_1, a_2, and a_3.	 Halite (cube) Pyrite Fluorite Galena  Magnetite (octahedron)  Pyrite
 TETRAGONAL	Three mutually perpendicular axes, two of the same length ($a_1 = a_2$) and a third (c) of a length not equal to the other two. Fourfold axis of symmetry around c.	 Zircon  Zircon  Fluorite (twinned)
 HEXAGONAL	Three horizontal axes of the same length ($a_1 = a_2 = a_3$) and intersecting at 120°. The fourth axis (c) is perpendicular to the other three. Sixfold axis of symmetry around c.	 Apatite  Apatite
 TRIGONAL	Three horizontal axes of the same length ($a_1 = a_2 = a_3$) and intersecting at 120°. The fourth axis (c) is perpendicular to the other three. Threefold axis of symmetry around c.	 Quartz  Corundum  Calcite (flat rhomb)  Calcite (scalenohedron)  Calcite (steep rhomb)  Calcite (twinned)
 ORTHORHOMBIC	Three mutually perpendicular axes of different length. ($a \neq b \neq c$). Twofold axis of symmetry around a, b, and c.	 Topaz  Staurolite** (twinned)
 MONOCLINIC	Two mutually perpendicular axes (b and c) of any length. A third axis (a) at an oblique angle (β) to the plane of the other two. Twofold axis of symmetry around b.	 Orthoclase  Orthoclase (carlsbad twin)  Gypsum  Gypsum (twinned)
 TRICLINIC	Three axes at oblique angles (α, β, and γ), all of unequal length. No rotational symmetry.	 Plagioclase

Figure 1.4

Characteristics of the seven crystal systems and some examples.

*(Copyright © 1974 McGraw-Hill Book Co. (UK) Ltd. From Cox, Price & Harte: An Introduction to the Practical Study of Crystals, Minerals and Rocks, Revised Edition. Reproduced by permission. Colors have been added to the original and are not accurate. They are shown for illustrative purposes only.) *Most laboratory collections of minerals for individual student use do not include crystals of these minerals. The collection may, however, contain incomplete single crystals, fragments of single crystals, or aggregates of crystals of one or more minerals. The best examples of these and other crystals may be seen on display in most mineralogical museums. **Staurolite is actually monoclinic but is also classified as pseudo-orthorhombic. Pseudo-orthorhombic means that staurolite appears to be orthorhombic because the angle β in the monoclinic system (see left-hand column under monoclinic) is so close to 90 degrees that in hand specimens it is not possible to discern that the angle β for staurolite is actually 89 degrees, 57 minutes.*

Figure 1.5 Quartz crystal.

Figure 1.6 Rose quartz.

Figure 1.7 Smoky quartz.

Figure 1.8 Chalcedony (chert).

Figure 1.9 K-feldspar.

Figure 1.10 Plagioclase.

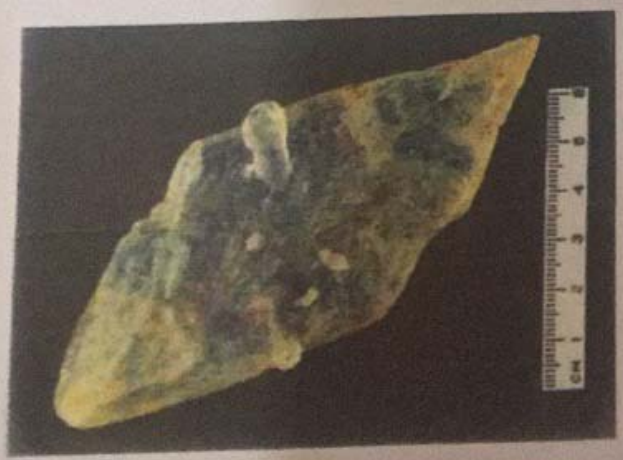


Figure 1.11 Gypsum (selenite).

Figure 1.12 Talc.

Figure 1.13 Calcite (note double refraction).

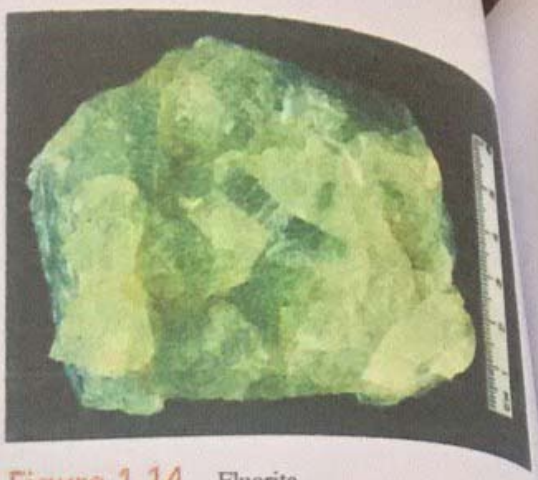


Figure 1.14 Fluorite.

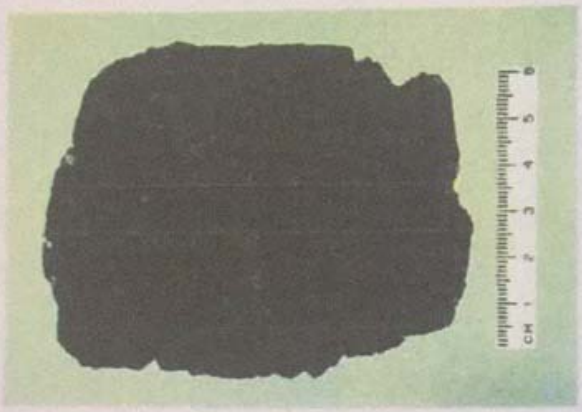


Figure 1.15 Biotite.

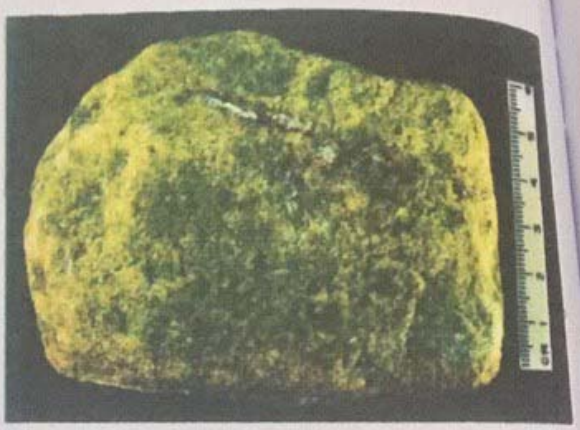


Figure 1.16 Olivine.

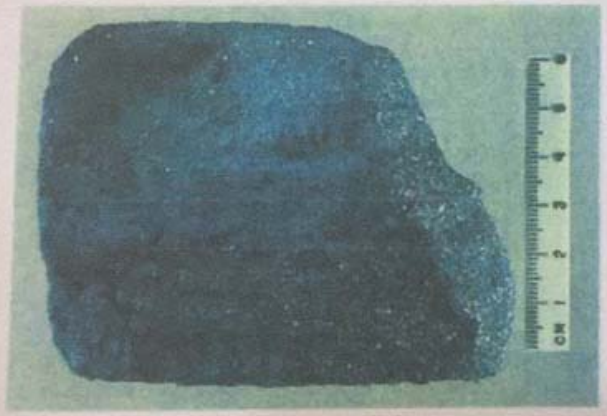


Figure 1.17 Hematite (specularite).

Figure 1.18 Hematite.

Figure 1.25 Pink granite.

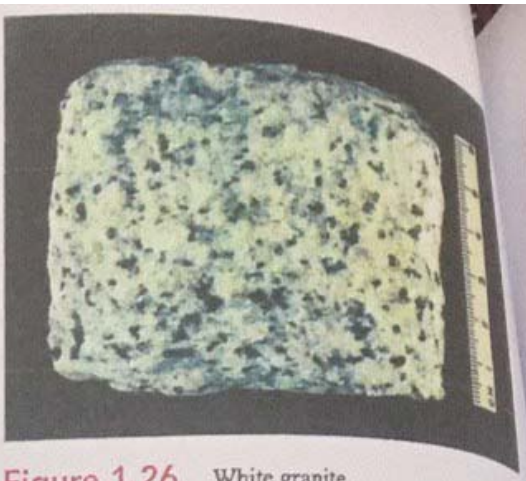


Figure 1.26 White granite.

Figure 1.27 Pegmatite.

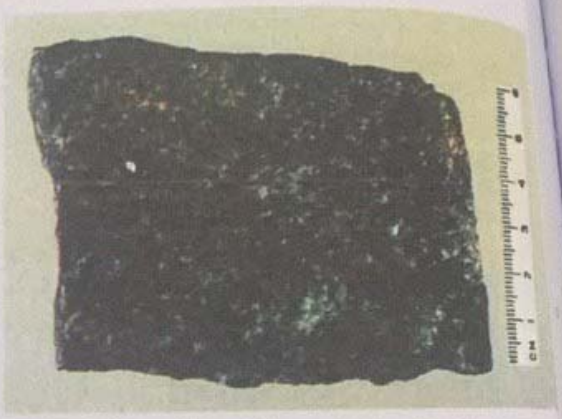


Figure 1.28 Gabbro.

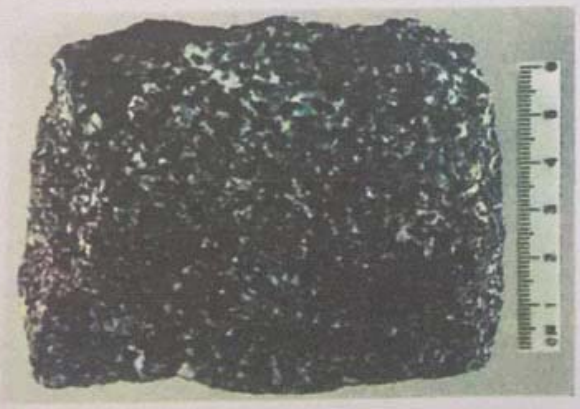


Figure 1.29 Diorite.

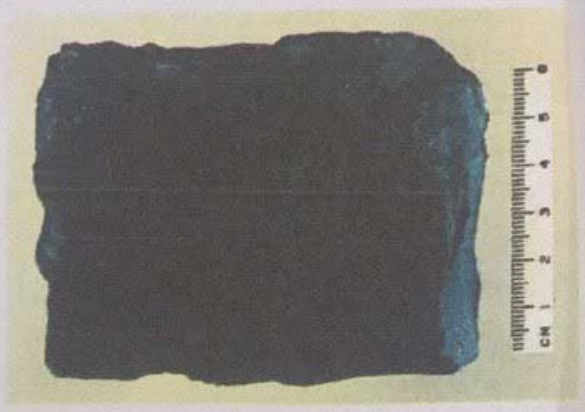


Figure 1.30 Basalt.

Figure 1.31 Vesicular basalt.

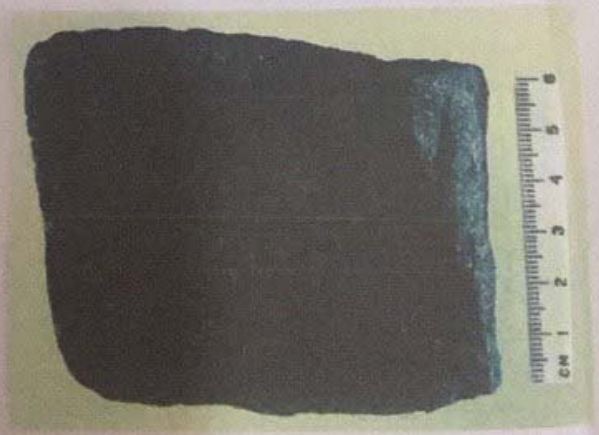


Figure 1.32 Basalt porphyry.

Figure 1.33 Rhyolite.

Figure 1.35 Rhyolite tuff.

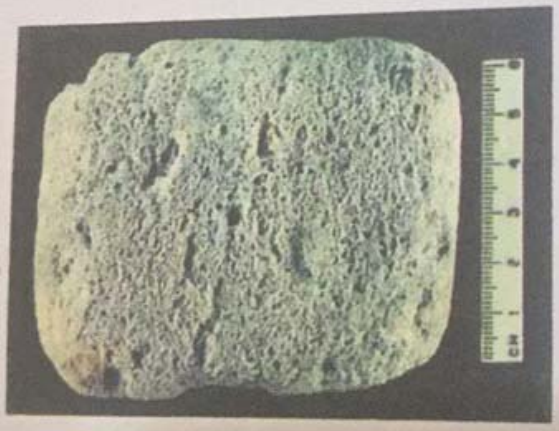


Figure 1.37 Pumice.

Figure 1.39 Obsidian.

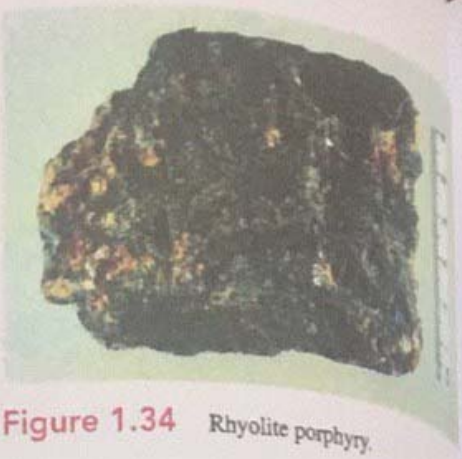


Figure 1.34 Rhyolite porphyry.

Figure 1.36 Andesite porphyry.

Figure 1.38 Scoria.

Figure 1.40 Peridotite.