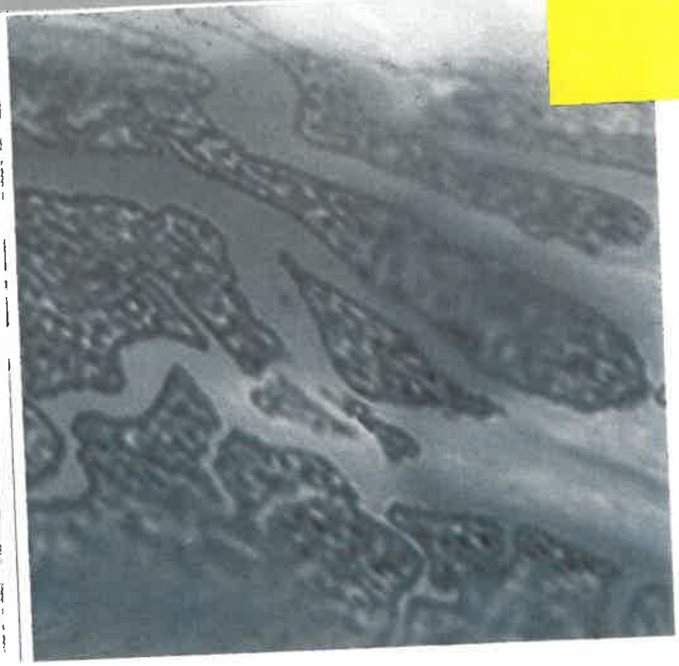


CHAPTER 7

Circulation of the Solid Earth

Plate Tectonics

Key Questions

- How do the physical and chemical characteristics of Earth change with depth toward its center?
- What is plate tectonics?
- What provides the energy that drives plate tectonics?
- How can we relate the surface features of Earth to plate tectonics?
- What is the rock cycle?
- How have the geographical positions of the continents changed through time as a result of plate tectonic activity?

Chapter Overview

Our understanding of solid Earth processes has taken a great leap forward since the mid-1960s with the development of the theory of *plate tectonics*. Yet there is still much to learn about the composition and dynamics of the solid Earth. Earthquakes and volcanoes demonstrate that the interior of Earth is not a static place. Rather, like the oceans and atmosphere clinging to its surface, the solid Earth is in motion. The energy that drives the circulation of the solid Earth derives not from the Sun but from Earth's interior. Convective currents in the interior are coupled to the rigid rocks that form the continents and seafloor, putting the continents in motion. Where continents collide, huge mountain belts form; where oceanic blocks collide with each other or with continents, deep-sea trenches and volcanoes form. These plate tectonic forces join with the surface processes of rock weathering and erosion to generate landscapes and recycle elements from solid Earth reservoirs into the soils, hydrosphere, and atmosphere, making those elements available to the biota once again. Thus, plate tectonic activity is critical to the maintenance of a biologically active planet.

We will study the circulation of the solid Earth in the way we studied atmospheric and oceanic circulation. First, we explore the anatomy of the planet, from its exterior to its greatest depths. The tools used to reveal Earth's internal structure give us clues about the temperature and compositional variations in the interior, but very little direct information exists. Then we discuss how the heat flux from the interior produces the motions within the solid Earth and how these motions form and modify Earth's major surface features: mid-ocean ridges and mountains, deep-sea trenches, and transform faults. We then trace the history of these motions over the past 3 billion years, during which time the continents have drifted apart and joined together again and again in a global tectonic cycle.

INTRODUCTION

The German meteorologist Alfred Wegener is largely credited with establishing the fundamental underpinnings of the theory that we now call **plate tectonics**. According to this theory, Earth's surface is divided into rigid plates of continent and ocean floor that move relative to each other through time. (*Tectonics* is the study of Earth's crust and the processes that deform it.) Wegener was fascinated by the near-perfect fit between the coastlines of Africa and South America and by the correspondence among the geological features, fossils, and evidence of glaciers on these two separate continents. Could all the continents once have been assembled into a *supercontinent?* Wegener believed so, as had others before him. He, however, was the first to put together all the diverse evidence in support of that concept. He named the proposed landmass **Pangea**, meaning "all Earth." He proposed that Pangea began to break apart just after the beginning of the Mesozoic era, about 200 million years ago, and that the continents then slowly drifted into their current positions. This theory is called **continental drift**. Wegener's maps, produced in 1924, are remarkably similar to the best global paleogeographic reconstructions available today (Figure 7-1).

Although it was welcomed by paleontologists because of its consistency with the terrestrial fossil record, Wegener's theory of continental drift was not well received by the geophysicists of his day. The British scientist Sir Harold Jeffreys presented calculations in 1925 demonstrating that the continents could not possibly plow through the rigid seafloor, as the theory seemed to require. Other scientists were unconvinced because Wegener could not propose a physical mechanism for driving the motion of the continents. Indeed, many of Wegener's own calculations and proposed mechanisms were found to be in error and untenable.

The acceptance of Wegener's theory of continental drift awaited a better understanding of the structure and operation of the solid Earth. That understanding has come about largely as the result of geophysical exploration, which has revealed the complex anatomy of Earth's interior and has led to the theory of *seafloor spreading*.

ANATOMY OF EARTH

Seismic Probing of Earth's Interior

For centuries geologists have been probing Earth's surface, cataloging the variety of rocks exposed and studying the processes that have led to their formation. The study of material that has risen to the surface has given some insight into the chemical and *mineralogical* composition of the shallow interior, but virtually everything we know about Earth's deep interior has been derived by indirect methods. Paramount among these methods is the science of *seismology*, the study of earthquakes and related phenomena.

EARTHQUAKES An **earthquake** is the sudden release of stored energy as a result of rapid movement between two blocks of rock. This energy radiates away from the earthquake in the form of vibrations. The site of energy release, known as the *focus*, can be anywhere from very near Earth's surface to as deep as 700 km below. This area, Earth's uppermost shell, is rigid; when it deforms, it does so *elastically*. This means that the material recovers its shape

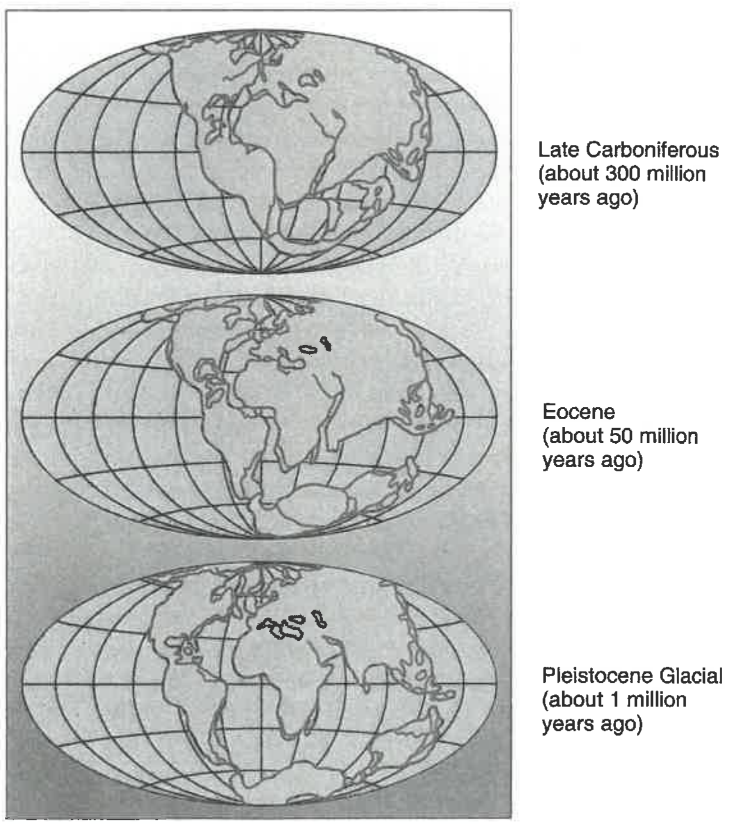


FIGURE 7-1 Wegener's reconstructions of the positions of the continents in the geological past. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems*, 8/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

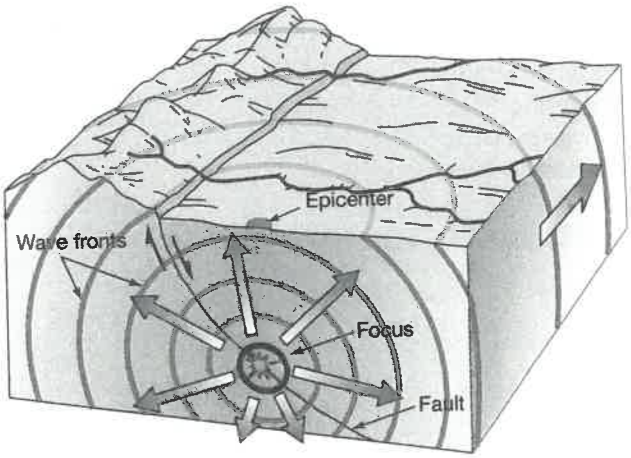


FIGURE 7-2 An earthquake's focus is located at depth. The point on the surface that directly overlies the focus is the epicenter. (*Source:* From J. P. Davidson, W. E. Reed, and P. M. Davis, *Exploring Earth: An Introduction to Physical Geology*, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

after the force that is tending to deform it is removed, unless it is deformed to the point of fracture and the original shape cannot be recovered. If there is differential movement on either side of the break, the fracture is called a *fault*. The *epicenter* of an earthquake is the position on Earth's surface directly above the focus (Figure 7-2).

Seismic Waves

Just as a person jumping into a swimming pool produces waves and ripples in the water, earthquakes create vibrations called **seismic waves** that ripple through Earth's interior, away from the earthquake's focus, as a result of that deformation. Two types of seismic waves are generated: body waves and surface waves. Both types spread outward from the focus. As we might expect, **body waves** travel through Earth's interior, whereas **surface waves** travel only across the surface. Surface waves transmit earthquake energy along Earth's surface, where movement is unconstrained vertically. The motion is much like that of a water wave, easily seen by watching a cork bobbing up and down: Particles are displaced upward, backward, downward, and then forward in a circular motion. There is no net movement of the particles, but energy is transmitted away from the center. Body waves are categorized as either P waves or S waves on the basis of their mode of propagation through Earth. **P waves**, or primary waves, result from the compression of material in Earth's interior. The material is alternately compressed and, as the wave travels away, stretched. Thus, a P wave travels as a series of compressions and expansions in the overall direction of wave movement, similar to the way sound travels or to the response of a spring or a Slinky® (Figure 7-3a). **S waves**, which are also called secondary or shear waves, are

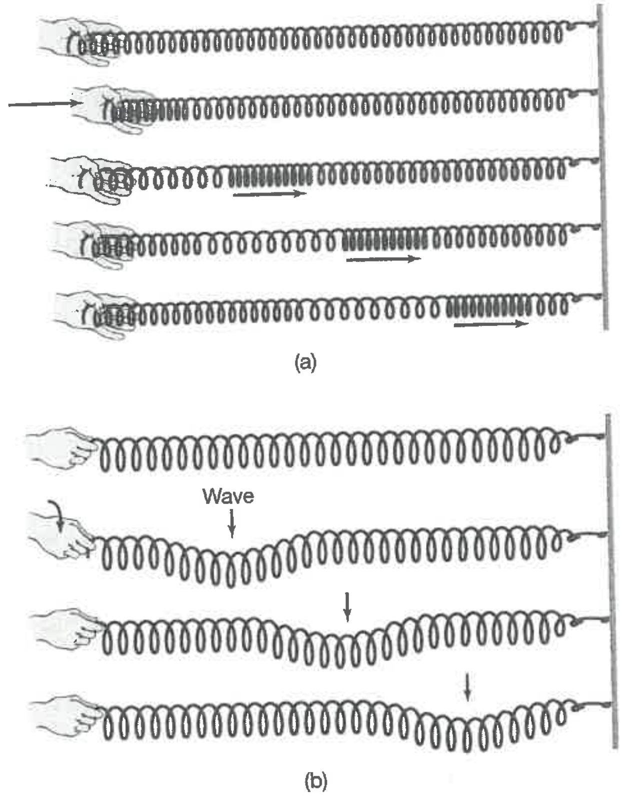


FIGURE 7-3 The movement of seismic body waves. (a) P waves alternately compress and expand materials, like the transfer of a compressive force imposed on a spring. (b) S waves move material from side to side, perpendicular to the overall direction of wave motion, just as a spring responds to repeated side-to-side shaking. (*Source:* From J. P. Davidson, W. E. Reed, and P. M. Davis, *Exploring Earth: An Introduction to Physical Geology*, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

transmitted as displacements perpendicular to the overall direction of wave travel. An analogy is the movement of a spring swung from side to side (Figure 7-3b). Although both solids and fluids can transmit P waves, only materials with structural rigidity (solids) can transmit S waves. This fact has proved to be important in the characterization of Earth's interior, allowing us to identify particular regions as fluids rather than solids, as we will see later.

Eventually the path of all body waves intersects Earth's surface. There they can be detected and recorded by a *seismograph*, which is a sensitive instrument that detects slight vertical and horizontal displacements of Earth's surface (see the Box "A Closer Look: The Principle of the Seismograph"). The rate at which seismic body waves travel through Earth depends on the properties of the material in Earth's interior. If we know how much time it takes waves to travel from the earthquake source to a site where they are detected at the surface by a seismograph, and if we can determine the path a particular seismic wave has taken, then we can calculate an average wave speed along that path. For a single earthquake event, a seismograph near the wave source records waves that traveled very shallowly through Earth, whereas a seismograph far from the source of the earthquake receives seismic waves that may have traveled through Earth's center. Thus, by comparing several seismographic records from various places around the world for a particular event, we can construct a fairly detailed three-dimensional view of the paths along which seismic waves travel through Earth. This process is called *seismic tomography* (Figure 7-4).

GENERALIZED STRUCTURE OF EARTH The gross picture revealed by seismic imaging is of a layered Earth comprising a *crust,* a *mantle* (consisting of an upper mantle and a lower mantle), an *outer core,* and an *inner core* (Figure 7-5), defined on the basis of contrasts in seismic wave velocities. The shallowest of these transitions is the crust–mantle boundary, first discovered by the Croatian seismologist Aa Mohorovičić, who was investigating shock waves traveling t hrough Earth from Zagreb (the former Yugoslavia) in the early 20th century. The boundary is now known as the *Mohorovčić discontinuity,* or the **Moho,** in his honor. This boundary is defined by a sharp increase in seismic wave velocities; P-wave velocities in the crust of around 5–6 km/sec increase to uppermost mantle velocities of about 8 km/sec. Beneath the continental crust, the depth to the Moho ranges from as much as 75 km in young mountain belts to 20 km in areas undergoing extension and crustal thinning. Beneath the oceanic crust, the Moho is at a nearly constant depth of around 7 km below the ocean floor.

Below the Moho, the velocities of both P waves and S waves generally increase with depth through the mantle, although a *low-velocity zone* (or LVZ) exists at a depth of between 80 and 300 km (see Figure 7-5). Velocities then increase again through the transition zone between the upper and lower mantles, but the increase is not smooth: It occurs in a stepwise fashion, which indicates some sort of incremental change in mantle properties. Seismologists think that this change is related to a transformation of the minerals present to more compact, denser forms. Seismic velocities then increase more gradually with depth through the lower mantle.

The boundary between the lower mantle and the outer core, at a depth of 2900 km, is distinguished by a significant drop in P-wave velocities and by the disappearance

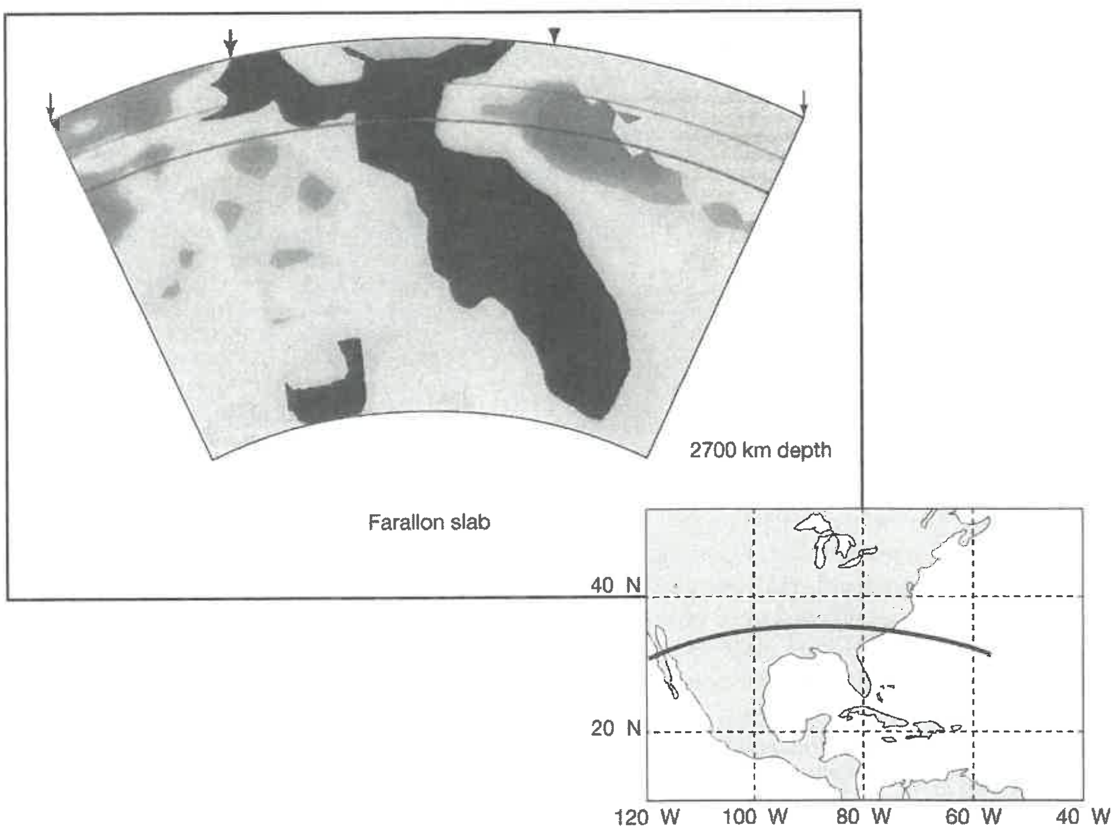
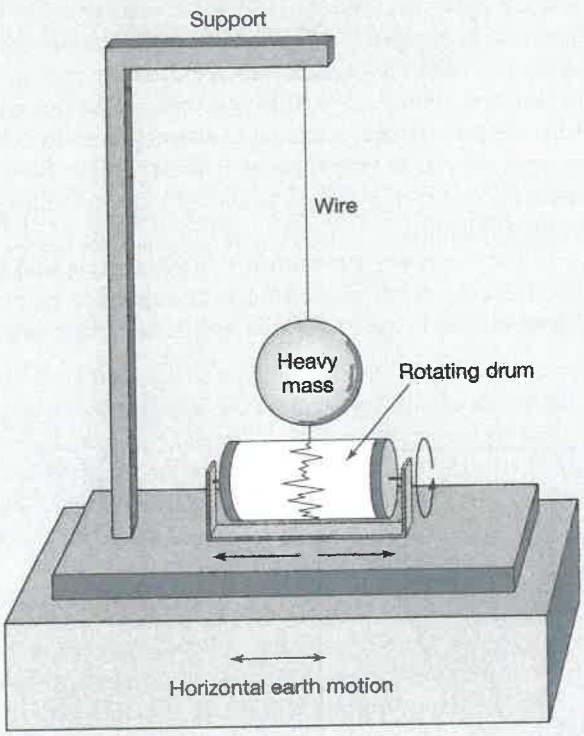


FIGURE 7-4 [See color section] Tomographic image of the mantle's S-wave velocity variations underneath North America, along the transect line shown in the insert. Blue colors indicate regions of fast seismic velocities, while reds indicate slow seismic velocities. The blue region cutting across the center of the diagram is the downgoing Farallon slab, which has been subducting under North America for 100 million years. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems,* 8/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

A CLOSER LOOK
The Principle of the Seismograph

In the simplest sense, seismographs consist of a recording drum, which is anchored to bedrock and situated below a suspended weight with a pen attached. The weight has sufficient inertia to remain vertical as the ground (and the recording drum) vibrate beneath it (Box Figure 7-1). During an earthquake, the recording drum vibrates back and forth, allowing the seismologist to determine the amount of displacement that occurs as the seismic wave passes through the ground below.

The *amplitude*, or size, of seismic waves is related to the amount of energy released during an earthquake. Earthquake amplitudes are reported according to the



BOX FIGURE 7-1 A seismograph. (*Source:* From S. Judson and S. M. Richardson, *Earth: An Introduction to Geologic Change*, 1995. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

(a)

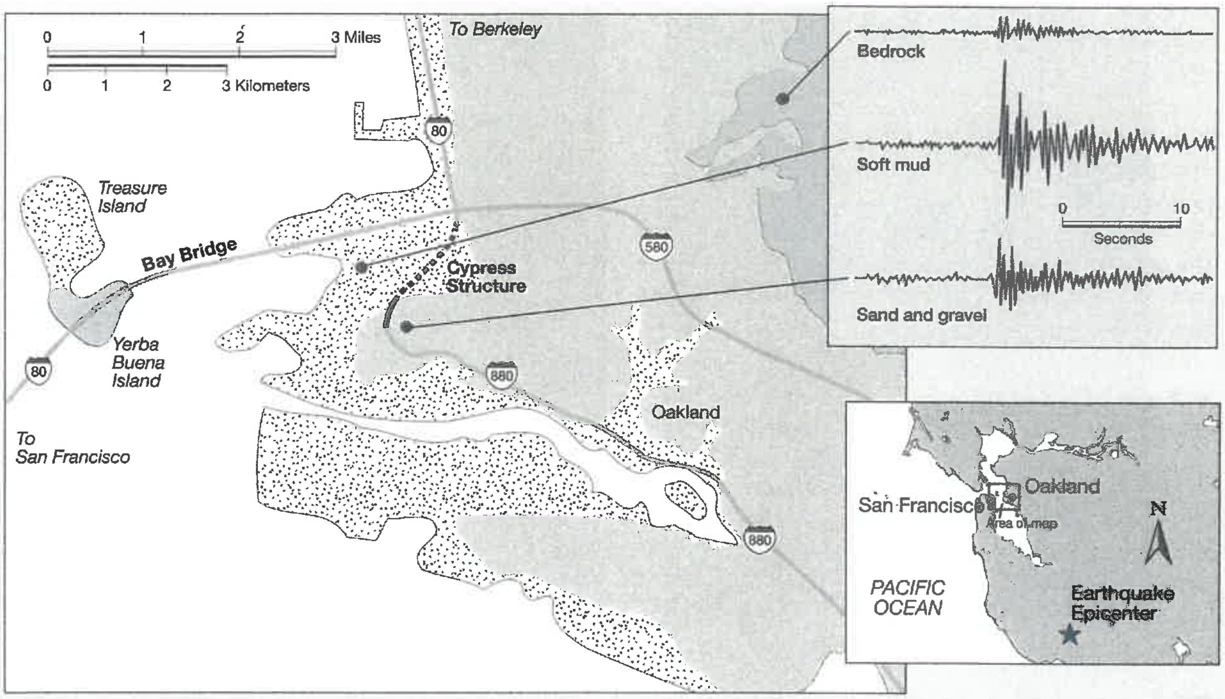
BOX FIGURE 7-2A Aerial view of collapsed section of the Cypress viaduct on Interstate 880, Oakland, California, after the 1989 Loma Prieta earthquake.

of S waves. Because S waves travel only through solids, geophysicists infer that the outer core is a metallic fluid. At a depth of 5150 km, P-wave velocities increase markedly. This increase in velocity measurably deflects P waves from their anticipated path, confirming that the inner core is solid. Tremendous pressure at Earth's center is thought to be responsible for converting fluids to solids there.

This description of Earth's interior structure, although essentially correct, is greatly simplified. Seismic tomography reveals a more heterogeneous distribution of seismic wave velocities (see Figure 7-4), perhaps related to the exchange of materials between Earth's interior and its surface. Clearly Earth's interior is a dynamic place. Plate tectonic activity has altered the chemical and thermal structure of the mantle and may have operated in response to influences conveyed through the mantle from the core. Let us explore these heterogeneities in more detail, utilizing information not only from seismology but also from *petrology*, the study of the origin and evolution of the chemical and mineralogical compositions of rocks.

The Crust

Earth's uppermost layer, the crust, is not homogeneous; rather, it varies in both thickness and composition. The most pronounced differences are between the continental and oceanic crust. *Continental crust* underlies the continents, whereas *oceanic crust* underlies the ocean basins. As delineated by variations in Moho depth, continental crust is thicker than oceanic crust. It is also less dense and



(b)

FIGURE BOX 7-2B Seismograms of the Loma Prieta earthquake recorded by seismographs situated on bedrock, sand and gravel, and soft mud. Notice how the amplitude and duration of ground shaking is greater in soft mud than in bedrock. (*Source:* S. Hough P. A. Friberg, R. Busby, E. F. Field, K. H. Jacob & R. D. Borcherdt, *Nature* 344, 1990, pp. 853–855; see also http://geopubs.wr.usgs.gov/fact-sheet/fs176-95.)

"moment magnitude" scale, an improved variant of the more familiar but now essentially obsolete "Richter" scale. For every tenfold increase in seismic wave amplitude (and approximately 30-fold increase in energy), the magnitude increases by one unit. Thus, an earthquake with a magnitude of 5 releases 30 times as much energy as one with a magnitude of 4.

Seismographs from around the world recorded the October 17, 1989, Loma Prieta earthquake with an epicenter 16 km northeast of Santa Cruz, California (Box Figure 7-2b). The earthquake had a moment magnitude of 6.9, creating the first major rupture along the San Andreas Fault since the famous 1906 San Francisco earthquake. Sixty-eight people died as a result of the Loma Prieta event, and nearly 4,000 were injured. Thousands of homes and businesses were damaged or destroyed, with an estimated dollar loss on the order of $7 billion. Structures built on mud and sand suffered larger amplitude and longer duration shaking than those on bedrock. One of the lasting images of the earthquake is the collapsed sections of the Cypress viaduct on Interstate 880 (Box Figure 7-2a).

on average older. The two types of crust differ in chemical and mineralogical compositions as well. To understand these differences, we must first introduce a genetic classification of crustal rocks.

IGNEOUS, SEDIMENTARY, AND METAMORPHIC ROCKS

All rocks are composed of **minerals**, defined as naturally occurring inorganic solids of definite crystal structure and chemical composition. Geologists recognize three major types of rocks: *igneous, sedimentary,* and *metamorphic.* **Igneous rocks** form by the cooling and solidification of **magma**, which is molten, or liquid, rock. If the magma solidifies beneath Earth's surface, the rocks are called *intrusive* igneous rocks. **Granite** is a well-known intrusive rock. If the magma is carried to Earth's surface at a volcano and erupts, it is called *lava* and cools rapidly, forming *extrusive* igneous rocks. **Basalt** is an abundant type of extrusive igneous rock. Both intrusive and extrusive igneous rocks vary in composition, especially in the amount of the mineral quartz (SiO_2) they contain. *Felsic* igneous rocks (e.g., granite or *rhyolite,* its extrusive analogue) are quartz-rich, light-colored, and less dense than *mafic* igneous rocks (such as basalt or *gabbro,* its intrusive analogue).

Rocks (of any type) that are exposed at Earth's surface tend to decompose, or *weather,* into finer materials called **sediments**—layers of unconsolidated mineral matter that is transported by water, wind, or gravity. As new sediments are deposited on top of existing sediments, the underlying sediments become compacted, expelling water from the pores between sediment grains. The remaining pores may become filled with mineral cements precipitated from subsurface fluids. Compaction and cementation

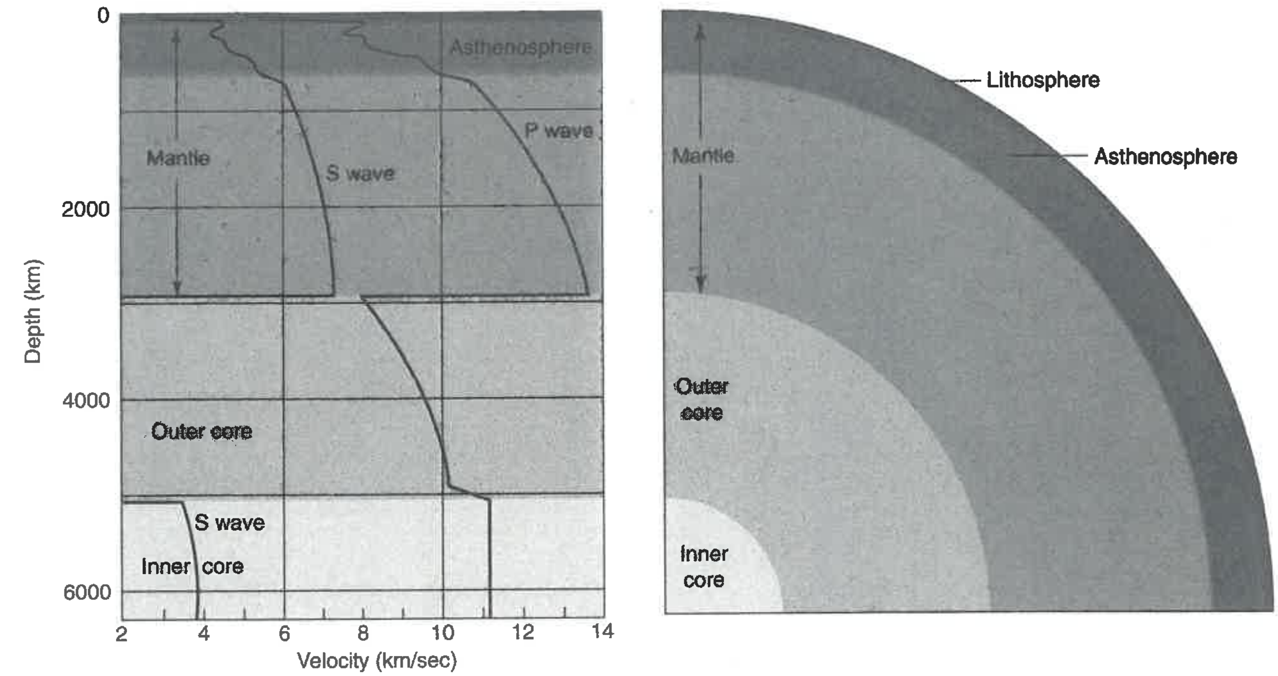


FIGURE 7-5 Internal structure of Earth, showing the distribution of seismic wave speeds. The speed changes with depth, defining the boundaries between the crust, mantle, outer core, and inner core. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems*, 9/e, 2001. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

are processes that contribute to **lithification**, the conversion of loose sediments into thick, cohesive, layered deposits known as **sedimentary rocks**. Sedimentary rocks formed from sand-sized grains (>63 μm in diameter) are called *sandstone*, whereas rocks composed of finer grains are called *mudstones*. Finely layered mudstones are commonly called *shales*. Sedimentary rocks may instead form chemically or biochemically. For example, some marine organisms form a shell or skeleton by precipitating calcium carbonate ($CaCO_3$) minerals. When the organisms die, the shells and skeletons accumulate on the seafloor and ultimately lithify into sedimentary rocks known as *limestones*.

Rocks (of any type) that are exposed to high temperatures, high pressures, chemically active fluids, or any combination of these agents are transformed in mineralogical and chemical compositions. As long as no melting is involved, the altered material is said to have been *metamorphosed* and is called a **metamorphic rock.** (When melting occurs, the resulting rock is igneous.) *Marble* is metamorphosed limestone, and *slate* is metamorphosed shale.

MAJOR ROCK-FORMING MINERALS Both continental and oceanic crust are composed primarily of rocks made of **silicate minerals**—that is, minerals rich in silicon and oxygen. *Feldspars* are the most abundant minerals in the continental crust. They are silicate minerals with aluminum, sodium, calcium, and potassium in their structures. Quartz also is an abundant silicate mineral in the upper continental crust. Together with certain other minerals, feldspars and quartz form granite and *granodiorite* (a slightly less quartz-rich rock), the hard, erosion-resistant rocks that create the high peaks of many mountain ranges, including the Sierra Nevadas of California.

Magnesium and iron-rich silicate minerals such as *olivine* and *pyroxene* characterize the basalts of the oceanic crust. The Hawaiian Islands are composed of basaltic materials. The high relative abundance of these dense minerals in the oceanic crust accounts for the higher density of mafic oceanic crust in relation to felsic continental crust.

SEDIMENTARY COVER Sediments and sedimentary rocks overlie the basalts and granites/diorites of the oceanic and continental crust, respectively. The source of these materials is easy to identify in the ocean: Sediments settle through the water column and accumulate on the seafloor as a sequence of relatively flat layers. Sedimentary rocks are also abundant in the continental crust, however. Some sedimentary rocks accumulated in basins on the continents themselves, but most were originally deposited as sediments on the seafloor and may have been deeply buried. Subsequent tectonic activity transported these sediments onto the continents. There they became exposed in mountain belts where the once-flat layering has typically become highly deformed through uplift. The oldest parts of the continents were once sedimentary rocks that have become significantly deformed and altered through many cycles of tectonic activity and metamorphism.

The Mantle

Beneath the crust is the mantle, which extends from the Moho to the top of the fluid outer core. The exact structure and composition of the mantle is a hotly debated topic among geologists, largely because the mantle is very difficult to observe. The samples of deeper mantle material that are available at the surface were brought up during rare geological events, such as the formation of kimberlites. *Kimberlites* are long, pipe-shaped igneous bodies that were emplaced after having passed rapidly from the upper mantle to the near-surface. They are remarkable in that they contain diamonds that formed under the high-pressure conditions of the mantle.

Most of what we know about the mantle is inferred from seismology. The velocity structure so determined indicates that the mantle is relatively uniform in composition and formed of silicate minerals. However, as depth increases so too do pressures and temperatures, causing changes in the structural and mineralogical composition of these silicates. These changes affect seismic wave velocities.

The recognition of the seismic low-velocity zone from depths of 80–300 km in the upper mantle proves to be an important link in plate tectonic theory. Most geologists accept that the low seismic wave velocities are the result of the presence of some molten rock at this depth. There need not be much; the data can be explained if only 1% or less of the rock is molten. Yet the small amount of melt present is critical, because it allows the crust and upper mantle to move relative to the underlying mantle—a basic tenet of plate tectonics.

A transition zone of rapidly increasing seismic wave velocities from depths of 370–650 km separates the upper and lower mantles. Geologists disagree on the reason for the transition zone. Many conclude from seismic evidence and theory that the transition zone is the consequence of mineralogical changes, whereas others conclude that differences in elemental composition are the cause. We return to this controversy later in our discussion of mantle convection.

The Core

The abundance of most elements in the crust and mantle can be explained by using meteorite compositions as a basis for comparison. Meteorites are thought to be fragments of larger *planetesimals* (bodies tens to hundreds of kilometers in diameter), many of which now reside in the asteroid belt of the solar system. Their parent bodies formed at the same time that the Sun did from the solar nebula. (This process is discussed further in Chapter 10.) A particular class of meteorites, the *carbonaceous chondrites*, is made of material that is thought to be essentially unaltered from the original nebular composition. Thus, carbonaceous chondrites are thought to be representative of the average abundances of elements in the solar system, including silicon, sodium, magnesium, calcium, and oxygen—all the basic rock-forming elements. Earth contains much less water and other *volatile* (easily vaporized) *compounds* as a consequence of its high-temperature formation, but we expect that its composition should otherwise be chondritic. By comparison with chondritic meteorites, the mantle and crust are significantly depleted in iron. This deficiency is made up for in the core, which is believed to be dominated by iron, along with small amounts (about 6%) of nickel, and approximately 8–10% of some unknown light element, which could be oxygen, sulfur, hydrogen, or silicon. (The light element has to be there; otherwise, the core would be even denser than it is observed to be.) The iron–nickel core is much denser than the overlying mantle, which, together with changes in seismic wave velocity, explains why seismic waves reflect off the core–mantle boundary.

Although the core is far removed from Earth's surface, it affects surface conditions because it is the source of Earth's magnetic field. Like a simple bar magnet, Earth has a magnetic field with north and south poles. Unlike a bar magnet, though, a magnetic dynamo (Figure 7-6) generates Earth's magnetic field. A

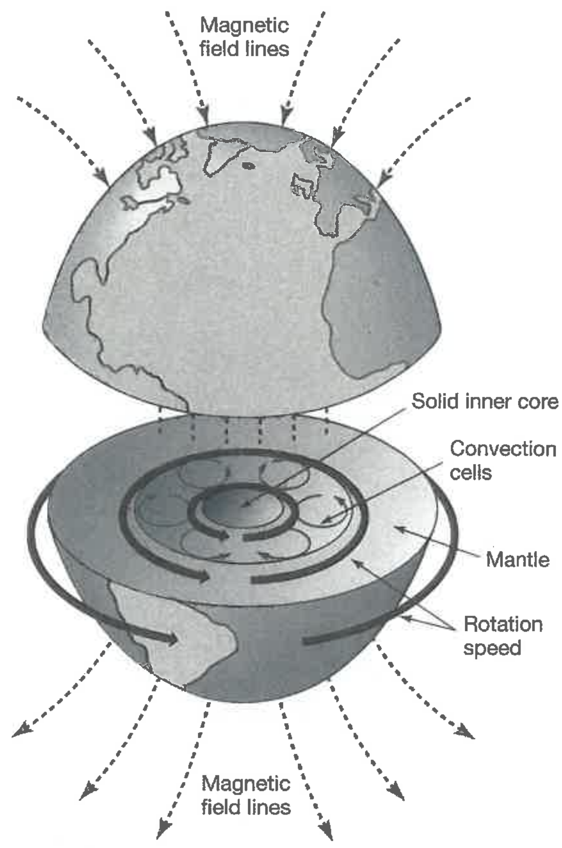


FIGURE 7-6 Earth's magnetic field is like that of a bar magnet, except that this field is generated electromagnetically by convection in the outer core. Dashed arrows indicate lines of force of the magnetic field. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems*, 8/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

magnetic dynamo is a mechanism that transforms energy from fluid motions (convection) into electrical currents that create a magnetic field. In a dynamo, the convecting fluid (liquid iron, in Earth's case) must be a conductor of electricity. The outer core convects because a temperature gradient is established across it by heat loss from the solid, inner core.

Why does the liquid outer core convect? As we discussed in Chapters 3 and 4, thermal convection occurs when a fluid is heated from below. Our example was the troposphere, which is heated by the warm Earth's surface. The ultimate source of this energy is sunlight. No sunlight makes it down to Earth's core, of course, so we must look for a different energy source there. One possibility is radioactive decay (which we cover later in this chapter), but the elements responsible for this heat source are preferentially concentrated in the crust and mantle rather than in the core. More likely, the energy required to drive convection in the outer core is derived from the gradual growth of the inner core. As Earth's interior cools, liquid iron slowly freezes out to form particles of solid iron. This "freezing" process releases heat, just as the freezing of water to form ice does. The newly formed particles of solid iron also heat the outer core frictionally as they settle down to join the inner core. The heat released by both of these processes is thought to be what powers outer core convection and, thus, the magnetic dynamo.

THE THEORY OF PLATE TECTONICS

Seafloor Spreading

As we discussed in the introduction to this chapter, the theory of continental drift lay more or less dormant for several decades after the publication of Wegener's ideas, largely because of the lack of acceptance by geophysicists. It is ironic that the resurgence of interest in Wegener's theory was the result of information obtained in the 1960s by geophysicists investigating the topographic and magnetic features of the seafloor. (*Topography* refers to the configuration of a surface, in particular the position and elevation of its features.) During and just after World War II, an intensive period of mapping took place that revealed intriguing details of the seafloor (Figure 7-7). This work gave the

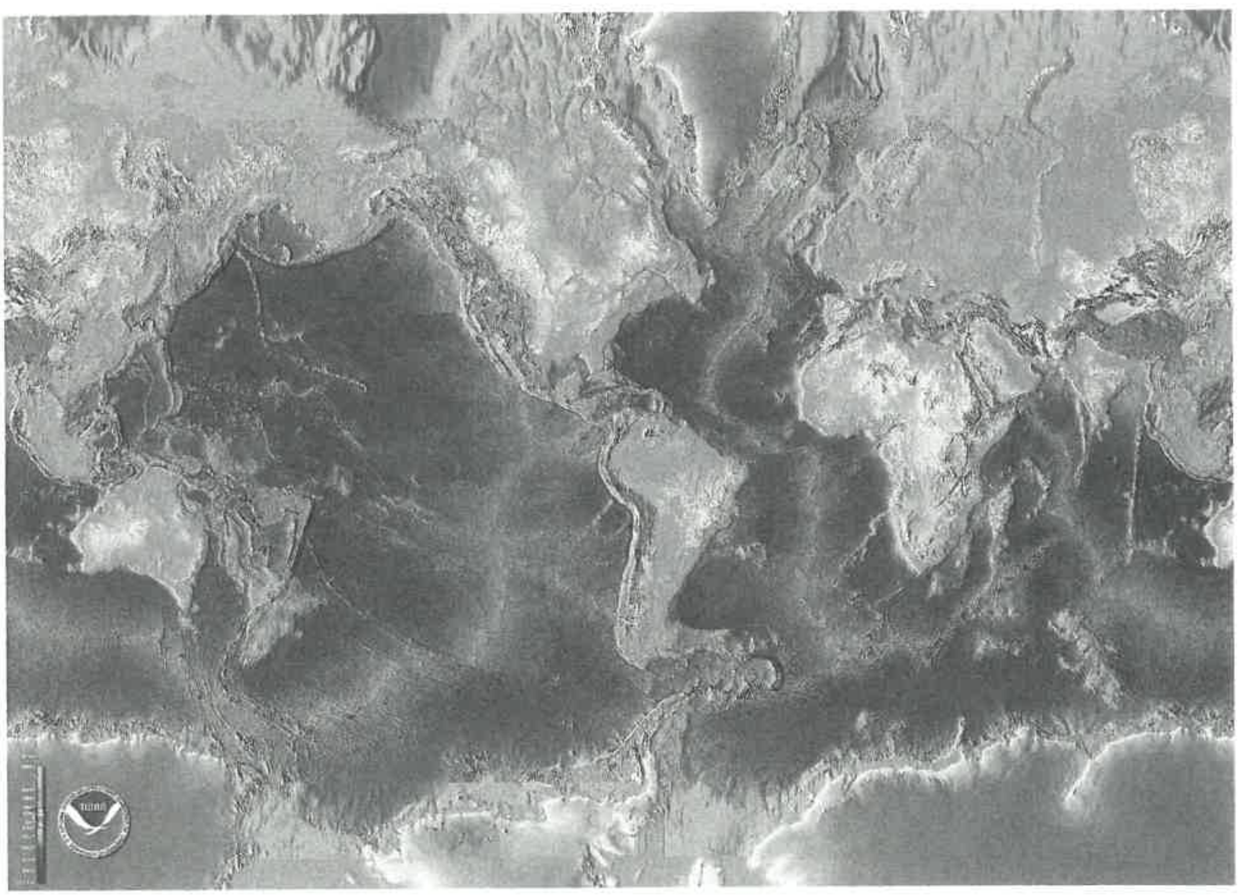


FIGURE 7-7 Ocean seafloor and continental topography inferred from high-resolution satellite altimetry measurements and ship depth soundings. Note transform faults extending across the Atlantic, especially near the equator and across the eastern Pacific. (*Source:* National Oceanic and Atmospheric Administration/Seattle.)

world the first evidence of chains of subsea volcanic mountains running down the centers of the ocean basins; we now call these features **mid-ocean ridges**. A *rift*, or narrow valley, runs down the center of such ridges. In the early 1960s, scientists proposed that these linear volcanic chains represent new seafloor that is extruded along the mid-ocean ridges. Once it forms, the new seafloor spreads to the sides of the ridges, generating the central rift, and is replaced at the *ridge axis*—the middle of the rift—by even younger new seafloor. This process was named **seafloor spreading**. Having just crystallized from the magma, newly formed seafloor at seafloor spreading centers is hot and expanded. As it spreads to either side of the plate boundary at the ridge axis, the material cools and contracts, and the seafloor subsides. This process occurs symmetrically across the axis of spreading and so creates symmetrical undersea mountain belts some 1000–4000 km wide that rise 2–3 km from the seafloor.

The real key to the origin of these features came from a better understanding of the magnetic characteristics of the seafloor. The *magnetic* **polarity** is the geographic orientation of the North and South poles. From studies of volcanic rocks extruded on land, scientists knew that magnetic polarity has flipped numerous times in Earth's history. The reasons are not well understood, but have to do with the complex behavior of the convecting liquid outer core. As the lava that formed these volcanic rocks cooled beyond a critical temperature of about 570°C (called the *Curie point*), the rocks became magnetized in the direction of Earth's magnetic field at the time of cooling. More-recent flows record switches in the magnetic field, with the North magnetic pole roughly coincident with the South geographic pole, and vice versa (Figure 7-8). Radiometric age dating (see below) provided the ages of volcanic flows.

Because the basaltic rocks of the seafloor were known to be of volcanic origin, in the late 1950s oceanographic expeditions were designed to map the magnetic character of the seafloor. As a result, a startling observation was made: The seafloor has a striped magnetic pattern, with the stripes running essentially parallel to the

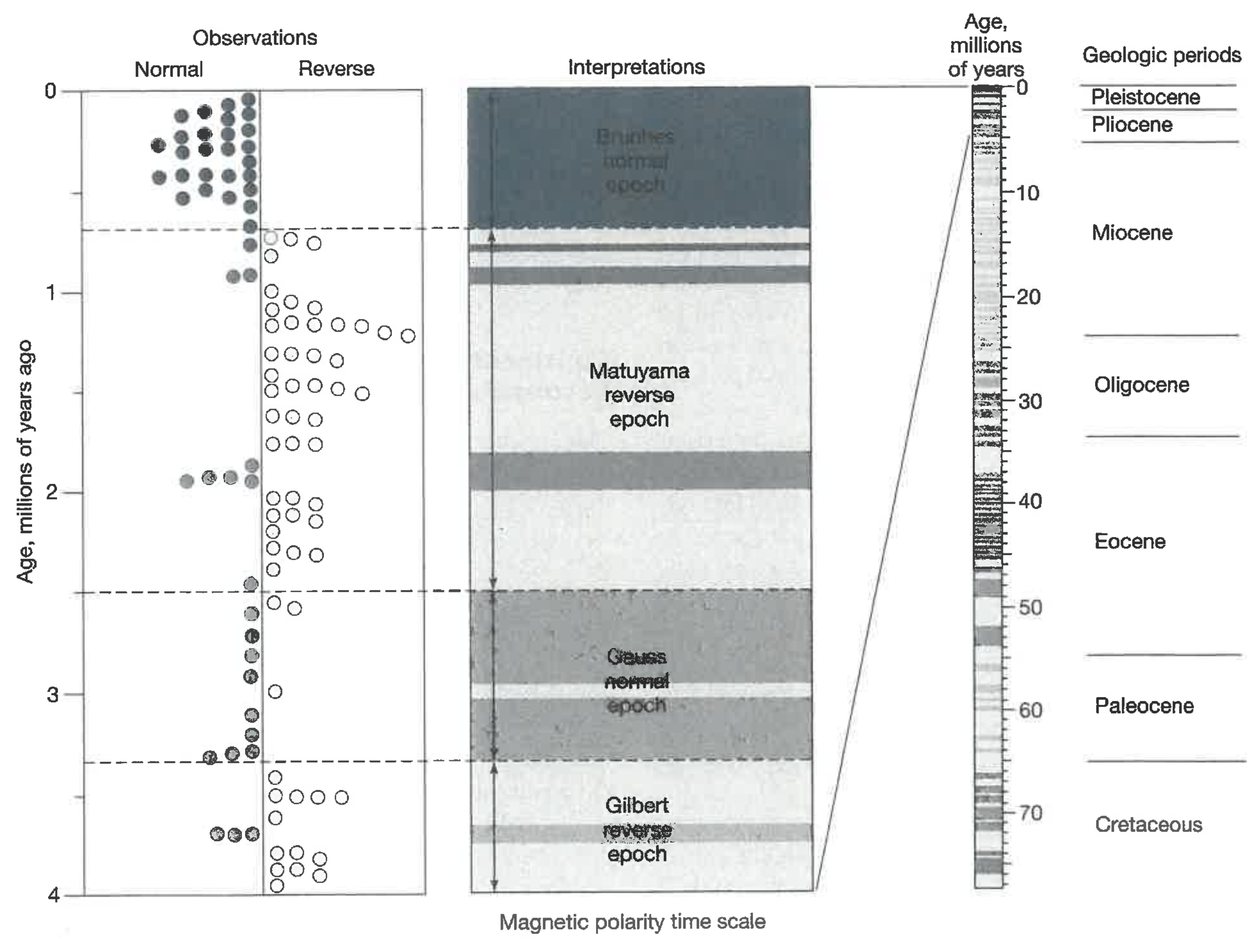


FIGURE 7-8 Magnetic reversals as recorded in volcanic rocks preserved on land for the last 75 million years. The last 4 million years of reversals are highlighted at the left. The pattern of change over a few million years is distinctive, and can be used as a signature for establishing the age of a sequence of rocks elsewhere in the world. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems*, 9/e, 2001. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

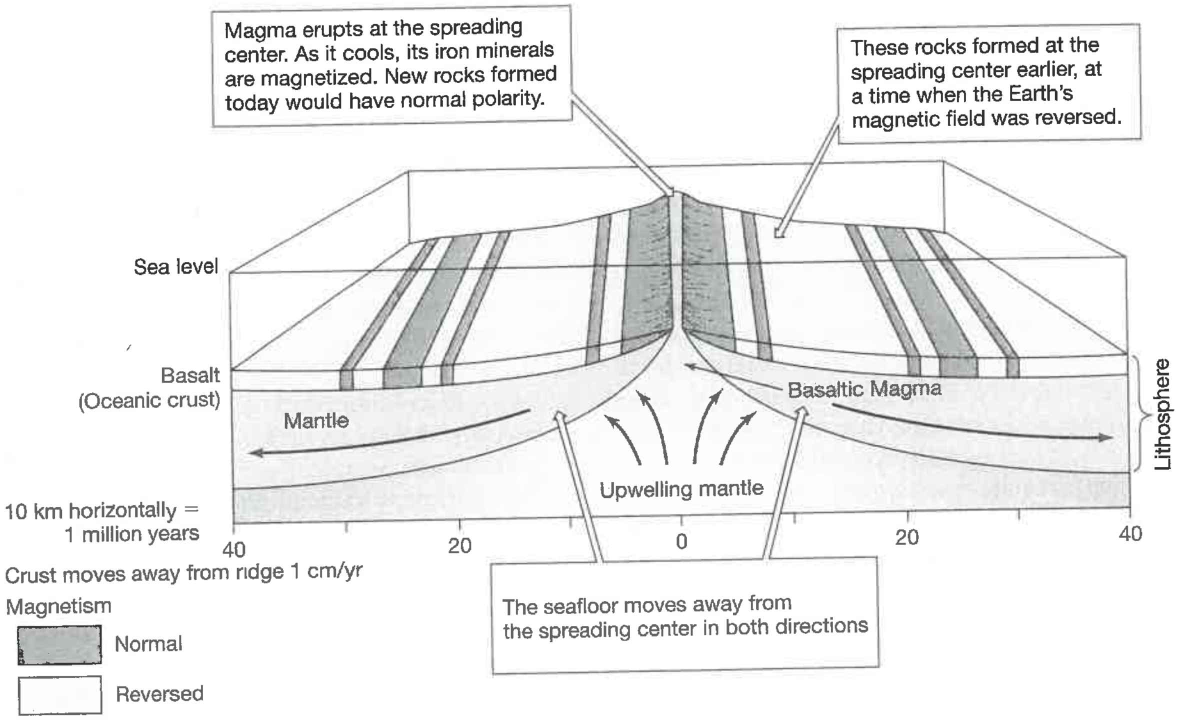


FIGURE 7-9 Magnetic stripes develop as new crust is added to the ocean floor at mid-ocean ridges and cools, becoming magnetized according to the magnetic field that exists at the time. As this material moves away from the axis, new seafloor is created, and its magnetization may be reversed if Earth's magnetization has reversed polarity in the intervening time. (*Source:* From S. Judson and S. M. Richardson, *Earth: An Introduction to Geologic Change*, 1995. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

mid-ocean ridges. The stripes reflect alternating bands of polarity. Today's magnetic polarity is considered to be *normal*, and the opposite polarity is considered to be *reversed* (Figure 7-8). Stripes on one side of a mid-ocean ridge were matched to others of similar width and polarity on the opposite side of the ridge (Figure 7-9).

Geophysicists soon concluded that this pattern of magnetic stripes must be caused by seafloor spreading. At the time of its formation, the new seafloor locks in the contemporaneous magnetic direction. It is then transported in opposite directions away from the ridge axis as new, molten rock is extruded from the volcano. Each reversal of the magnetic field produces a magnetic stripe on the seafloor. New material forming at a mid-ocean ridge is thereby differentiated from the older seafloor material that was produced during a previous magnetic interval at the same ridge and has subsequently drifted away from the ridge axis.

Seafloor spreading provided the solution to the problem plaguing geologists interested in continental drift since the days of Wegener: How could the continents drift through the rigid seafloor? The answer was that the continents do *not* plow through the seafloor. Rather, continents and segments of ocean floor are connected into plates that continuously move away from one another at mid-ocean ridges. Geologists have used a variety of types of evidence to reconstruct this drift of the continents throughout Earth's history, largely confirming the notion of Pangea proposed by Wegener decades earlier.

Continental Drift and Paleogeographic Reconstructions

The symmetrical magnetic stripes on the seafloor increase in age to either side of the Mid-Atlantic Ridge, recording the opening of the Atlantic Ocean (Figure 7-10). In a sense, we can reverse time by rolling the seafloor back into the Mid-Atlantic Ridge, bringing together once again stripes of equal age at the ridge axis. In this technique, the Atlantic Ocean closes first in the Southern Hemisphere and then in the Northern Hemisphere, as South America slips into place alongside Africa. The close fit between the two continents, especially at the edges of their continental shelves, was part of what convinced Wegener in the early 1900s that the continents have drifted apart over geological time.

Seafloor magnetic stripes provide the best tool for *paleogeography*, the reconstruction of the positions of the continents in the past. However, the tectonic process of *subduction* (see below) has destroyed much of the seafloor record of the past 200 million years and all of the record older than that. So, paleogeographers turn to other sorts of evidence to determine ancient continental positions.

Sedimentary rocks prove very useful in this regard. Glacial deposits generally form at high latitudes (poleward

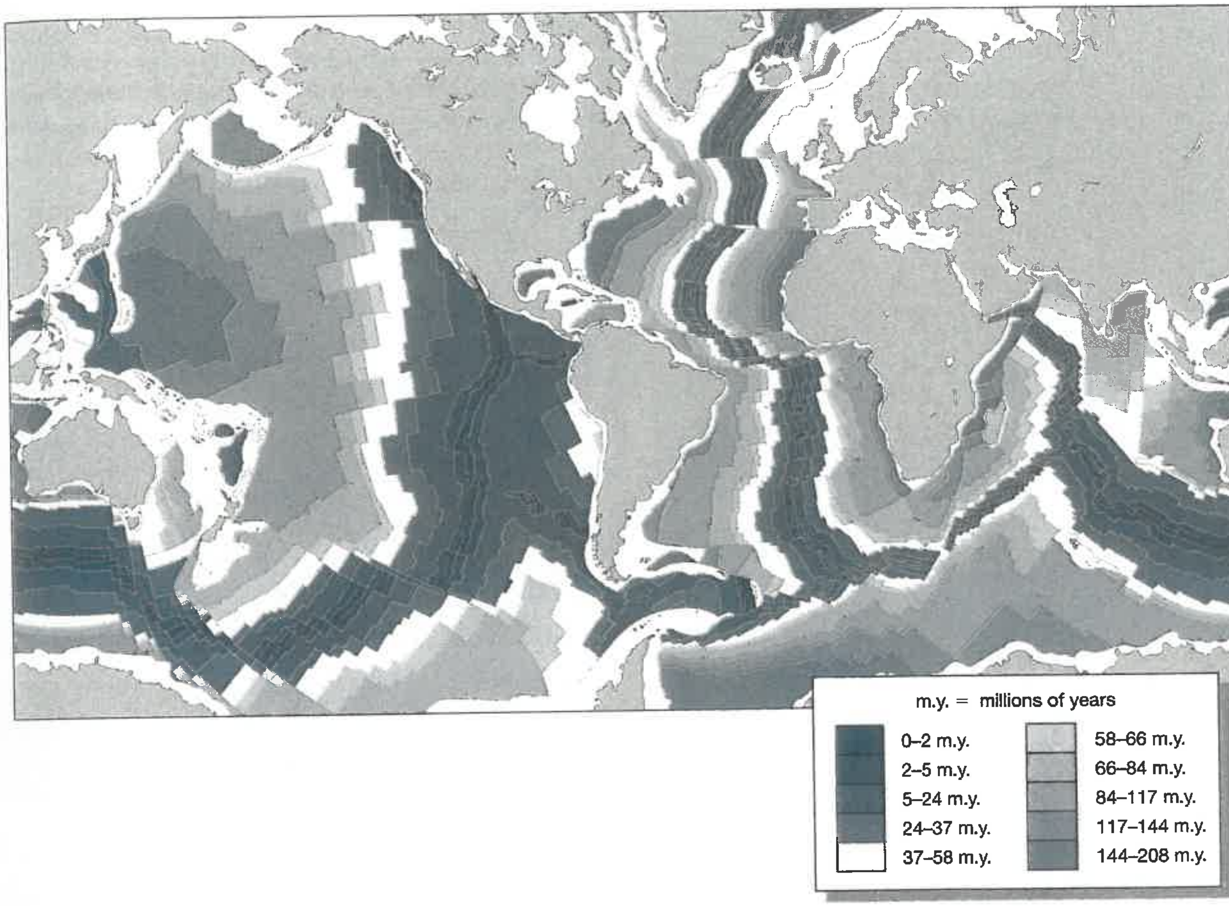


FIGURE 7-10 [See color section] The age of the ocean floor is shown as bands of different color on the basis of the magnetic striping developed during seafloor spreading. The youngest ocean floor is near the mid-ocean ridge, while the oldest is furthest away. (*Source:* From R. W. Christopherson, *Geosystems: An Introduction to Physical Geography*, 3/e, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

of about 45°), so we assume that glacial sediments indicate high *paleolatitudes*—that is, the latitudes at which the rocks formed. (We discuss an important exception to this rule in Chapter 12—the so-called Snowball Earth episodes of the Paleoproterozoic and Neoproterozoic, for which it appears that ice sheets extended into tropical latitudes.) Similarly, because coral reefs are located in the tropics today, we assume that reef limestones indicate tropical paleolatitudes. Salt deposits indicate subtropical paleolatitudes because they form preferentially in arid regions underlying the descending branches of the tropical Hadley cells. In addition, similar fossils on two continents indicate that the continents were in close proximity, or joined, at the time the organisms lived, allowing their migration. One must be cautious, however, in applying these paleolatitude indicators, because they are based on the assumption that the present is the key to the past. There may be times, however, when this assumption does not hold. Thus, geologists are motivated to look for more reliable paleolatitude indicators. The least ambiguous paleolatitude indicator comes from magnetism in rocks. By measuring the angle of the preserved magnetic field with respect to the sedimentary layering of the rocks (which is presumed to have originally been horizontal), geologists are able to estimate the angle of the magnetic field with respect to the horizontal. This angle gives the latitude at which the sedimentary deposit formed (Figure 7-11). Rocks in which the magnetic field is nearly parallel to the bedding plane must have formed near the equator; rocks in which the magnetic field is perpendicular to the bedding plane must have formed near the poles. These inferences, of course, are based on the assumption that Earth's magnetic field has always had two poles (North and South) and that the magnetic poles have always been approximately aligned with the geographic poles. One might rightfully point out that this is also a case of using the present to interpret the past. However, in this case, the assumption seems to be well-founded. During the past several million years when the continents could not have drifted very far, the magnetic poles have remained close to the geographic poles. Magnetic dynamo theory also suggests that, far away from its source (the outer core), the magnetic field should always be more or less aligned with the planet's spin axis.

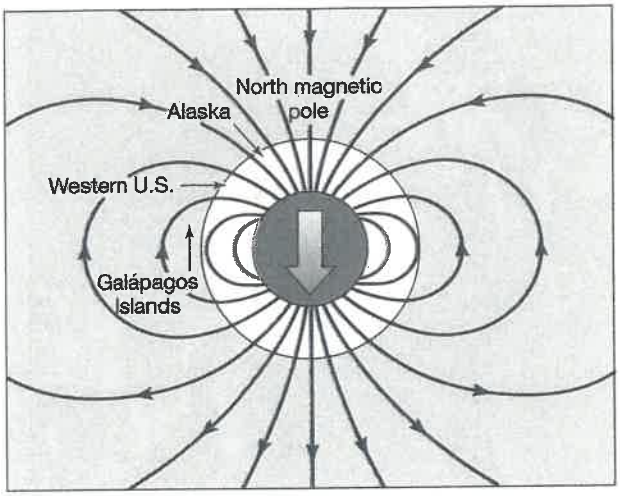


FIGURE 7-11 Earth's magnetic field, indicating that the angle of the field lines with respect to Earth's surface varies from horizontal at the equator to vertical at the poles. Paleogeographers use this feature to determine paleolatitudes of rocks, on the basis of the magnetic field orientation they acquired at the time of their formation. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems*, 8/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

Unfortunately, most of the tools paleogeographers use constrain only the latitude, not the longitude, of the continents. Longitudinal positioning is much more difficult. Similarity of fossil assemblages on adjacent continents is taken to indicate proximity, whereas large differences between the types of fossils found on two continents imply that wide oceans separated the continents and prevented the dispersal of organisms. However, features other than oceans, mountains and deserts for example, can also prevent dispersal. Thus, there are large uncertainties in the longitudinal positions of the continents prior to 200 million years ago.

The sequence of maps shown in Figure 7-12 shows how the continents are thought to have drifted over the past 500 million years, based on all available evidence. In the Cambrian and Ordovician periods (540–440 million years ago), the continents became increasingly dispersed from equatorial to southern high latitudes. Over the next 300 million years, the continents drifted together and collided. The collisions created large, Himalayan-style mountain ranges, including the Appalachians of the eastern United States. By 280 million years ago, the continents were assembled into the gigantic supercontinent Pangea, centered on the equator. Pangea eventually began to disassemble about 200 million years ago. The Atlantic Ocean formed as a rifting apart of continents first between North America and Africa, next between South America and Africa, and finally between North America and Europe. By 120 million years ago, Africa, Antarctica, India, and Australia had begun their separate paths. North America rotated and drifted from the low latitudes into its present position. Most impressive was the long journey of India as it detached from Antarctica and ultimately collided with Asia some 50 million years ago. We will see later on (Chapter 12) that this ongoing collision between India and Asia may be responsible, at least in part, for our present, relatively cool global climate.

The paleomagnetic and geological evidence of continental positions before 600 million years ago is sparse.

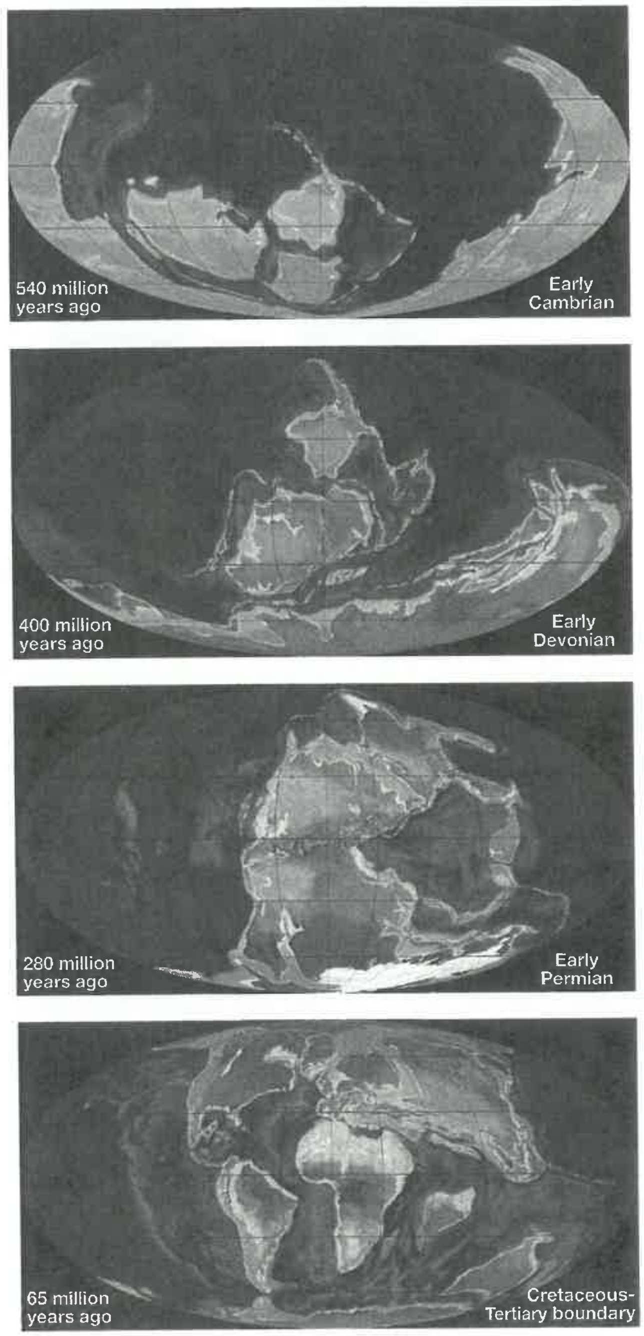


FIGURE 7-12 [See color section] Paleogeographic reconstructions from the Early Cambrian to the Cretaceous-Tertiary boundary. (*Source:* From R. Blakey, *used with permission.)*

Available paleomagnetic data seem to indicate that between about 900 and 600 million years ago the continents were assembled into another Pangea-like supercontinent. Prior to this time the data are so scarce that paleogeographic reconstructions are not currently feasible.

New Structural Categories: Lithosphere and Asthenosphere

As Wegener learned long ago, the conventional separation of the solid Earth into the core, the mantle, and the crust on the basis of seismic wave velocities is inadequate in view of plate tectonic theory. To explain the drift of continents, the mobile plates need to be distinguished from the lubricating layer below. To do so, the mantle and crust are best recategorized according to material strength (Figure 7-13). The plates extend through the crust and into the uppermost mantle; we call this outermost sphere the **lithosphere.** The upper, crustal part of the lithosphere is *brittle*, that is, it fractures in response to stress. Below the lithosphere is the **asthenosphere,** a region of the upper mantle that acts more like a fluid than a solid. The asthenosphere is *ductile*—it flows plastically, or deforms easily, in response to stress. The top of the asthenosphere is coincident with the mantle's low-velocity zone. Recall that the preferred explanation for low seismic wave velocities there is the presence of a small amount of molten rock. The asthenosphere extends through the mantle's transition zone to a depth of around 700 km. Below this depth, the lower mantle is thought to be much less ductile because of the effects of very high pressure.

PLATES AND PLATE BOUNDARIES

According to the theory of plate tectonics, the lithosphere is divided into about 20 rigid plates (Figure 7-14). The crustal portions of some plates are entirely oceanic, whereas other plates include both oceanic and continental crusts. *Oceanic lithosphere* describes a plate that is topped by oceanic crust; *continental lithosphere* refers to a portion of a plate topped by continental crust.

We now know that tectonic activity (such as earthquakes or volcanism) is concentrated at plate boundaries; there is little activity within a plate. This activity is the result of plate motion: The plates move relative to each other at average speeds of a few centimeters per year. As a result of friction between the plates, there are alternating periods of stasis (during which stresses build) and periods of movement (when they are released) both at the plate boundary and near the surface. (Seismic and satellite measurements indicate that at greater depths or farther from the plate boundary, the motions are more continuous.) After a period of stasis, pentup energy is released suddenly as the plates jump past each other, causing earthquakes. As predicted, the distribution of earthquakes at Earth's surface follows plate boundaries quite closely (compare Figures 7-14 and 7-15).

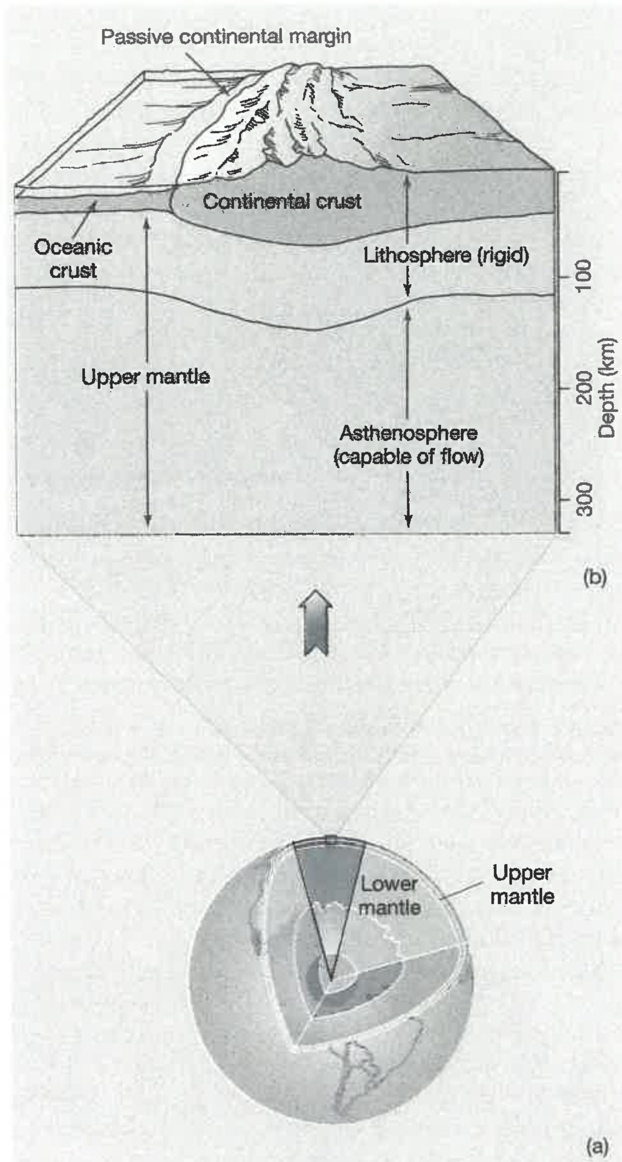


FIGURE 7-13 (a) Internal structure of Earth, comparing the traditional classification by seismic wave velocities with the plate tectonic classification by material strength. (*Source:* From R. W. Christopherson, *Geosystems: An Introduction to Physical Geography*, 3/e, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.) (b) A cross section of the upper mantle and crust showing the relative positions of the lithosphere (crust plus uppermost mantle) and asthenosphere. (*Source:* From J. P. Davidson, W. E. Reed, and P. M. Davis, *Exploring Earth: An Introduction to Physical Geology*, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

There are three types of plate boundaries (or margins): divergent, convergent, and transform (Figure 7-16). At *divergent margins,* lithospheric plates are moving away from each other. At *convergent margins,* plates are moving toward each other. At *transform margins,* plates are slipping past each other. Each boundary type is represented differently at Earth's surface. In other words, each type of

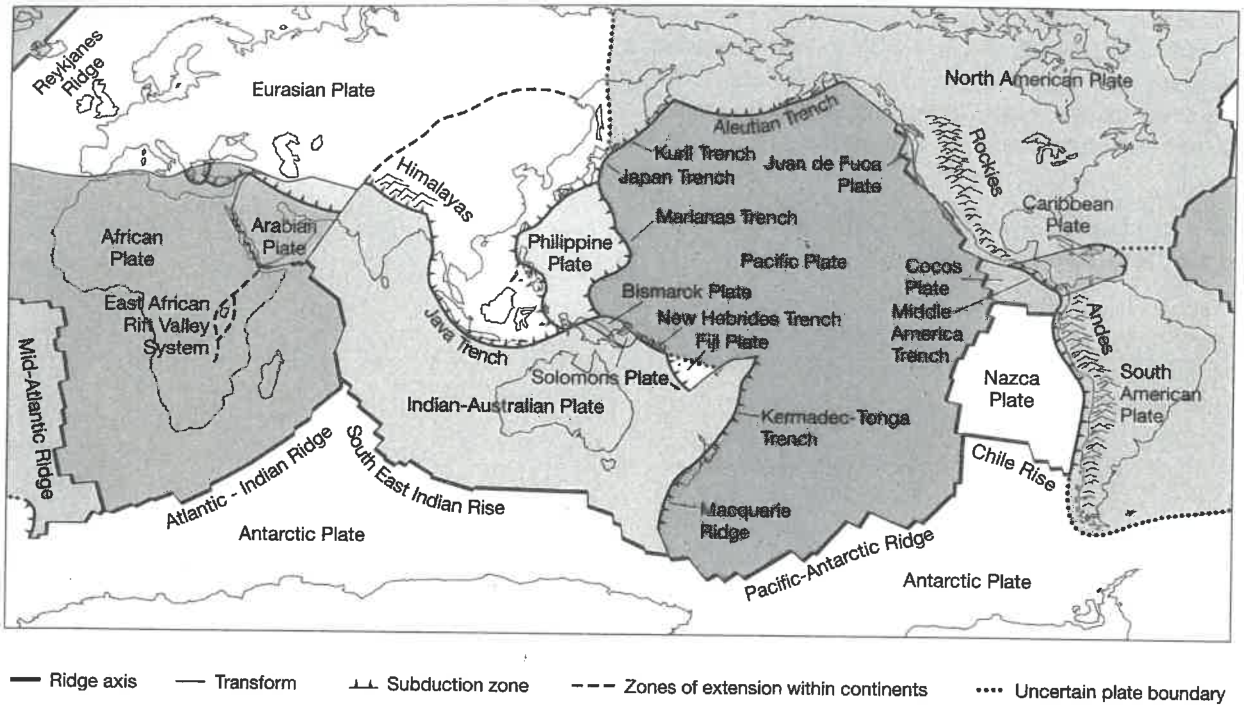


FIGURE 7-14 The lithosphere is divided into rigid plates. (*Source:* From S. Judson and S. M. Richardson, *Earth: An Introduction to Geologic Change*, 1995. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

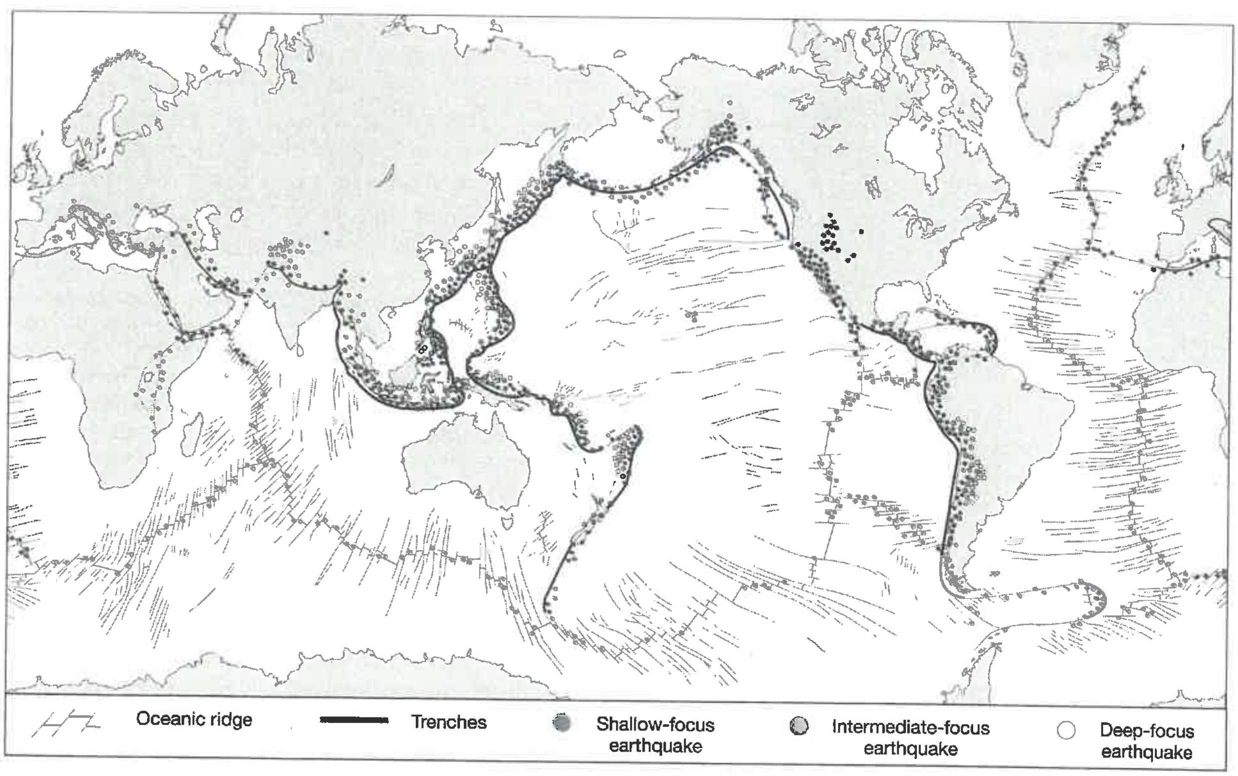


FIGURE 7-15 [See color section] Distribution of earthquakes of shallow, intermediate, or deep focus. Deep-focus earthquakes occur only at subduction zones. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems*, 8/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

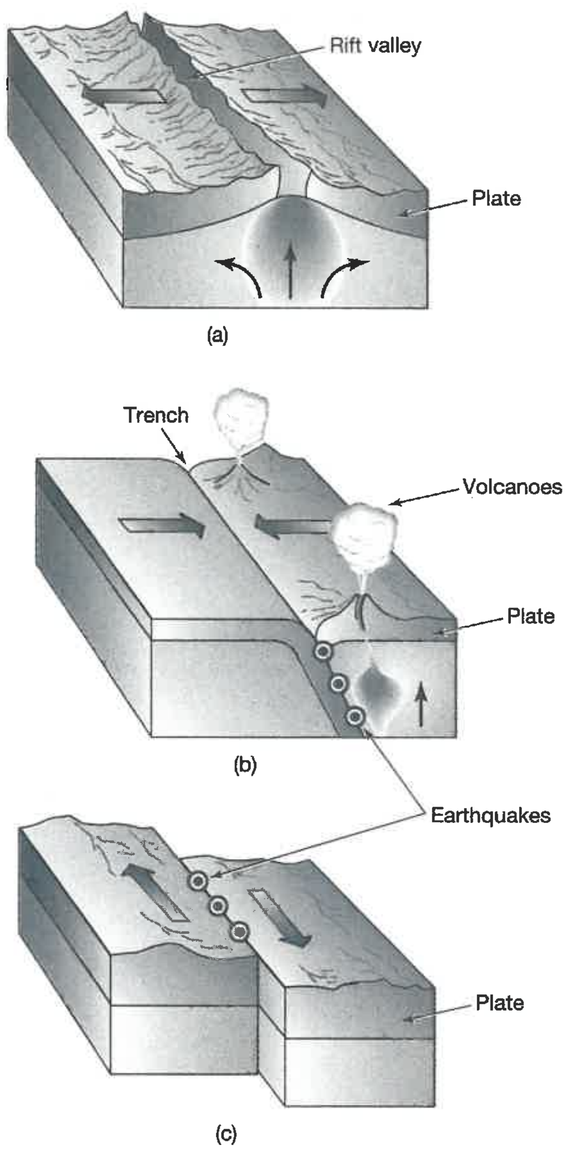


FIGURE 7-16 The three types of plate boundaries: (a) divergent; (b) convergent; and (c) transform fault. *(Source:* From J. P. Davidson, W. E. Reed, and P. M. Davis, *Exploring Earth: An Introduction to Physical Geology,* 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

plate margin is reflected in distinctive surface features: mid-ocean ridges, deep-sea trenches, and transform faults, respectively.

Divergent Margins

Divergent margins are regions where stresses are pulling apart the lithosphere. The mid-ocean ridges, already described, represent most of the divergent plate boundaries on Earth. Divergent boundaries that occur on land represent sites of continental fragmentation, or *rifting,* where the continental crust stretches. Tensional forces pull the continent apart. In the process, faulting occurs and flat-bottomed valleys called *rift valleys* form. During the breakup of Pangea some 200 million years ago, these types of plate boundaries were much more common than they are today. Today the rift valleys of East Africa, along with the Gulf of Aden and Red Sea to the north, are our best example of a divergent continental boundary in the making (Figure 7-17). If one studies this area in more detail, one finds that the divergence of the African and Arabian plates has created a system of rifts that radiate from a central point. This point is referred to as a *triple junction.* The East African rift valleys are at an earlier stage of rifting than the Red Sea and Gulf of Aden, where spreading has progressed to the point that new ocean basins have been formed.

Convergent Margins

Convergent margins are regions where two lithospheric plates are forced together. Although there has been some spirited controversy in the past, most geologists now are convinced that Earth is not increasing in size. (Like Wegener, those who have suggested that Earth changes its size have no physical explanation for how this could happen. Wegener, though, turned out to be right!) But if new seafloor is produced at mid-ocean ridges, then what happens to the old seafloor? The mapping efforts that followed World War II revealed deep basins in addition to mid-ocean ridges. Deep-sea trenches are long, narrow, very deep basins that are especially common along the margins of the Pacific Ocean (Figure 7-18). The discovery of these trenches provided the answer: The seafloor is consumed at deep-sea trenches about as fast as it is being produced at mid-ocean ridges.

Deep-sea trenches form at two of the three types of convergent plate boundaries: those that involve two oceanic plates and those that involve an oceanic and a continental plate. A third type of convergent margin forms the world's highest mountains rather than the deepest trenches; these involve the collision of two continental plates. Let us explore each of these cases.

OCEANIC–CONTINENTAL AND OCEANIC–OCEANIC CONVERGENT MARGINS Recall that the upper portion of a lithospheric plate (the crustal section) is brittle. When oceanic plates collide, the leading edge of one plate sinks entirely beneath the other. When the leading edge of one of the plates at a convergent margin is oceanic lithosphere and the leading edge of the other plate is continental lithosphere, the denser oceanic lithosphere sinks beneath the less-dense continental plate (Figure 7-19a). The sinking of an oceanic plate at a convergent margin is called **subduction**, and the entire region is called a *subduction zone.* The downgoing plate, called a **slab**, subsides into the mantle. In so doing, the plate bends, creating deep, linear depressions at the surface—deep-sea trenches. These trenches are the deepest parts of the oceans. Friction between the downgoing plate and the overriding plate generates

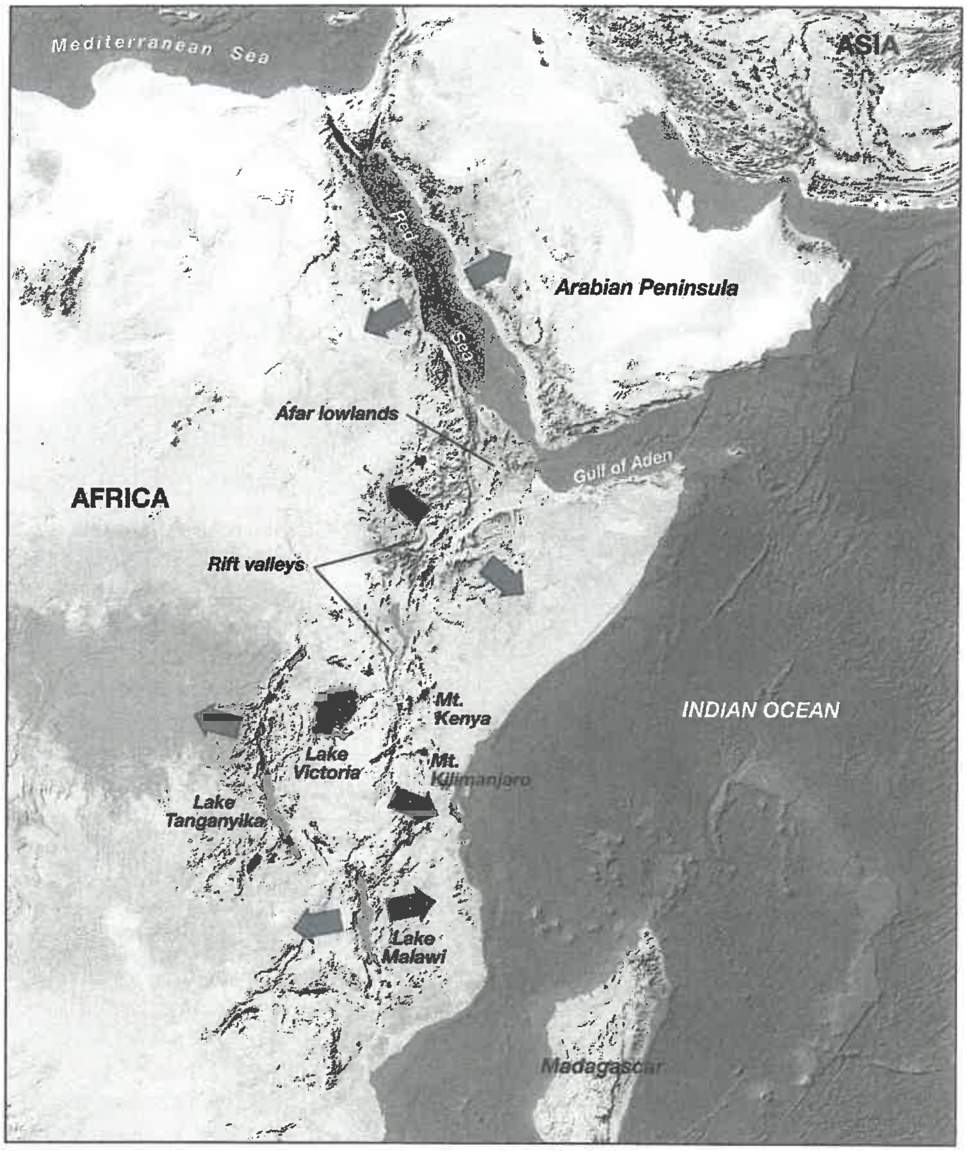


FIGURE 7-17 The rift valleys of East Africa, where Africa is being uparched and pulled apart. If spreading continues, the rift system may evolve into an elongate sea like the Red Sea to the north.

a substantial amount of seismic activity near the surface (in the upper 60–100 km); other forces generate earthquakes within the subducting slab at greater depths. The earthquake foci deepen as the distance from the trench toward the continent increases (Figure 7-15). Inland from the trench, water released into the mantle from the heated subducted slab of lithosphere leads to melting and produces igneous activity at the surface. This activity forms a range of volcanic mountains called a *volcanic arc.*

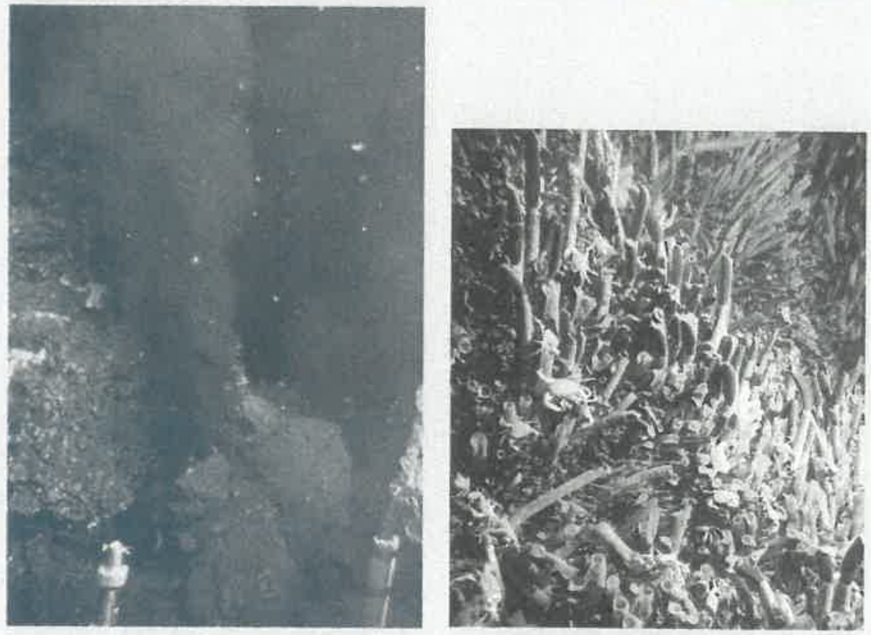
When two oceanic plates collide at a convergent margin, one subducts beneath the other (Figure 7-19b). Similar to what occurs in an oceanic–continental collision, a range of volcanic mountains forms to one side of the trench, but in this case the volcanoes rise up along the seafloor rather than on land. If they reach the ocean surface, they produce volcanic *island arcs.* The Marianas Islands, off the coast of the Philippine Islands, formed in this way. The Marianas Trench is the deepest trench of all, more than 10.5 km below sea level.

As the seafloor spreads from its place of origin at a mid-ocean ridge to its place of destruction at a subduction zone, sediment settling through the overlying water accumulates on the seafloor. Like a conveyor belt, the convergent motion of the plates in the subduction zone carries this sediment toward the trench. There the sediment may be scraped off by the opposing plate, forming wedges of deformed sediment. The rest of the sediment remains attached to the oceanic plate. The fate of this sediment is an area of active research. Part of it appears to become *underplated*—that is, attached to the base of the overlying plate—whereas some of it is carried into the asthenosphere. In the asthenosphere, it undergoes *dehydration* (loss of water) and *decarbonation* (loss of carbon) to the surrounding mantle, as well as a host of mineralogical

A CLOSER LOOK

Deep-Sea Life at Mid-Ocean Ridge Vents

The axial portion of a mid-ocean ridge is marked by volcanic activity, earthquakes with a shallow focus, and the venting of hot fluids rich in dissolved metals and hydrogen sulfide. Seawater drawn into the ridge along its flanks flows through cracks in the oceanic crust and is expelled through vents in the axis. Along the way, the seawater is heated, and chemical reactions with basalt alter its composition: Magnesium is removed and sulfate is reduced to sulfide, and calcium and trace metals are added. The circulation of seawater through the mid-ocean ridge also alters the chemical composition of the oceans. While exiting through the vents, iron sulfide minerals precipitate from the altered seawater and give the plumes a black coloration. For this reason, the vents through which the fluids exit are called *black smokers* (Figure Box 7-3a). They are also known as *hydrothermal vents*, because they release heated seawater.



(a) (b)

BOX FIGURE 7-3 [See color section] **Abundant and bizarre life thriving under the harsh conditions of the deep-seafloor, in the vicinity of hydrothermal venting. (a) A black smoker chimney is shown spewing out sulfide-rich solutions that provide the energy source for this food chain. (b) Tube worms, crabs, and other organisms can be seen. (*Source:* (a) Dudley Foster/Woods Hole Oceanographic Institution and (b) American Geophysical Union.)**

The fluids released by black smokers sustain a unique community of organisms. These organisms synthesize organic matter with the help of bacteria that carry out **chemosynthesis** rather than photosynthesis. In other words, these bacteria utilize energy from inorganic matter—energy released during chemical reactions between seawater and hydrogen sulfide. Chemosynthetic bacteria do not use the energy of sunlight. Feeding off these bacteria are unusual species of clams, crabs, and giant red and white tube worms (Box Figure 7-3b).

transformations. These reactions prove to be very important to the global recycling of elements such as carbon, because volcanoes that form in such subduction zones derive carbon dioxide gas from this sedimentary source.

CONTINENTAL–CONTINENTAL CONVERGENT MARGINS

When two continental plates meet at a convergent margin, the continents collide abruptly. Because continental crust is too buoyant to be subducted, continental collision results in the separation of the crustal portion of the lithospheric plate from the mantle portion below. Subduction of the mantle portion of one plate may occur while the continental crust on both plates becomes compressed and crumpled. As a consequence, tall mountain belts and high plateaus form (Figure 7-19c). Around 50 million years ago, India collided with Asia. The collision between these two segments of continental lithosphere led to massive deformation and uplift of the continents. The tall peaks of the Himalayas and the uniformly high Tibetan Plateau bear firm testament to the awesome energetics of this collision. Older mountain belts (such as the Appalachians) are the products of collisional tectonics that occurred hundreds of millions of years ago. Subsequent erosion has reduced what were once majestic mountains into the more modest ridges we observe today.

Transform Margins

When the relative motion along a plate boundary is parallel to the boundary, lithosphere is neither created (extruded) nor destroyed (subducted); the plates merely slip past one

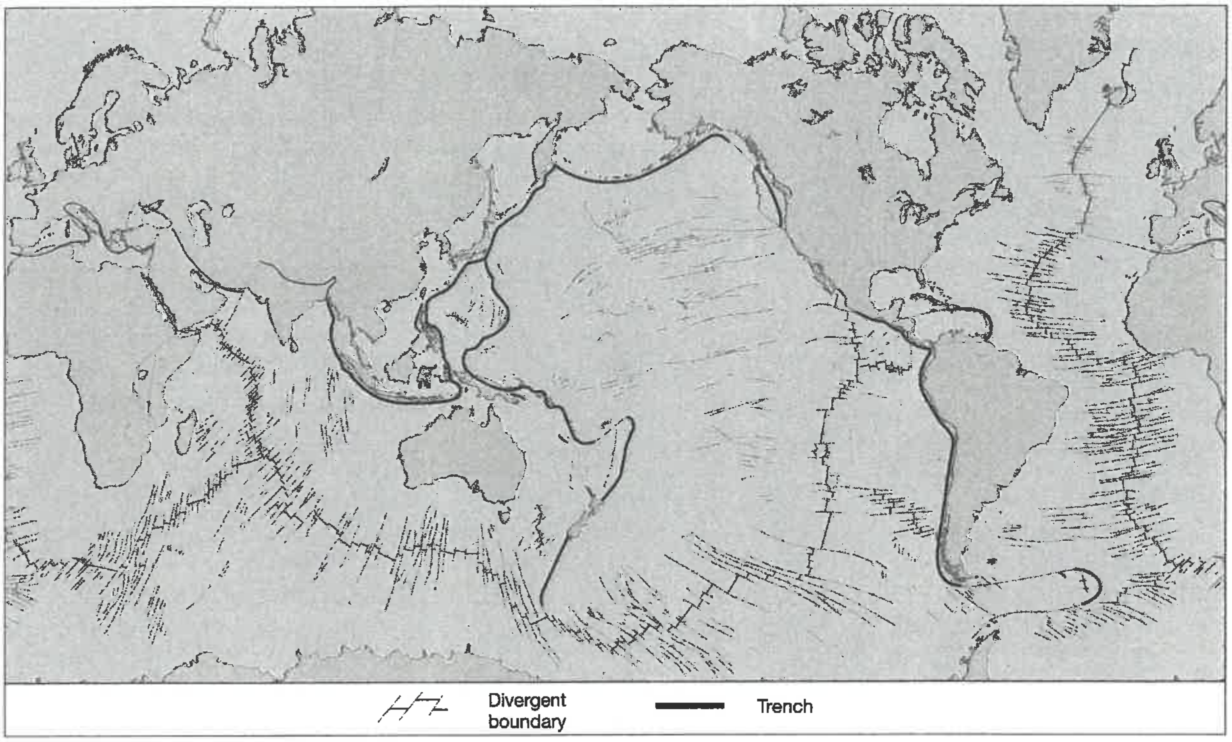


FIGURE 7-18 Distribution of oceanic trenches and mid-ocean ridges. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems*, 8/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

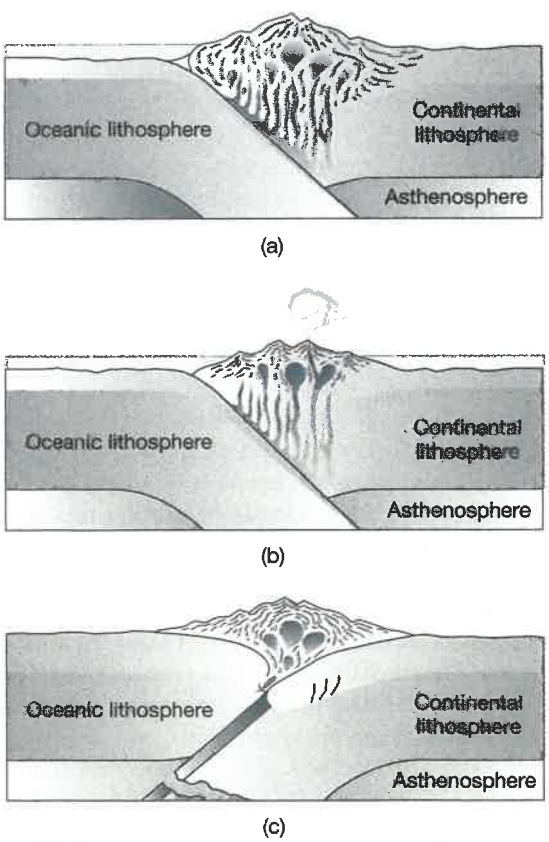


FIGURE 7-19 Three types of convergent plate boundaries: (a) oceanic–continental; (b) oceanic–oceanic; and (c) continental–continental. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems*, 8/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

another at a fault. Faults that form boundary-parallel margins are known as **transform faults.** The San Andreas Fault of California, which marks a segment of the boundary between the North American and Pacific plates, is a transform fault (Figure 7-20). Baja California and southern California (including Los Angeles) are moving slowly northward relative to the rest of California. In 50 million years or so, San Francisco and Los Angeles will be side by side, and beyond that time Los Angeles will actually be north of San Francisco. Geologists on the West Coast joke that California politics will at this time become completely reversed, with the northern part of the state being more conservative than the south.

Transform plate boundaries occur in oceanic settings as well. The jagged shape of parts of the mid-ocean ridge system is caused by offsets between ridge segments created by transform faulting (Figure 7-18). The Mid-Atlantic Ridge shows this type of behavior near the equator.

Overview of Plate Interactions

Figure 7-21 provides an overview of the types of plate interactions and the surface features they generate. The production of new oceanic lithosphere at mid-ocean ridges is matched by the destruction of older oceanic lithosphere at subduction zones, which are manifested at the surface by deep-sea trenches. Part of the sedimentary layer riding atop the oceanic plate is incorporated into the overriding plate (oceanic or continental), and the rest is subducted into the mantle.

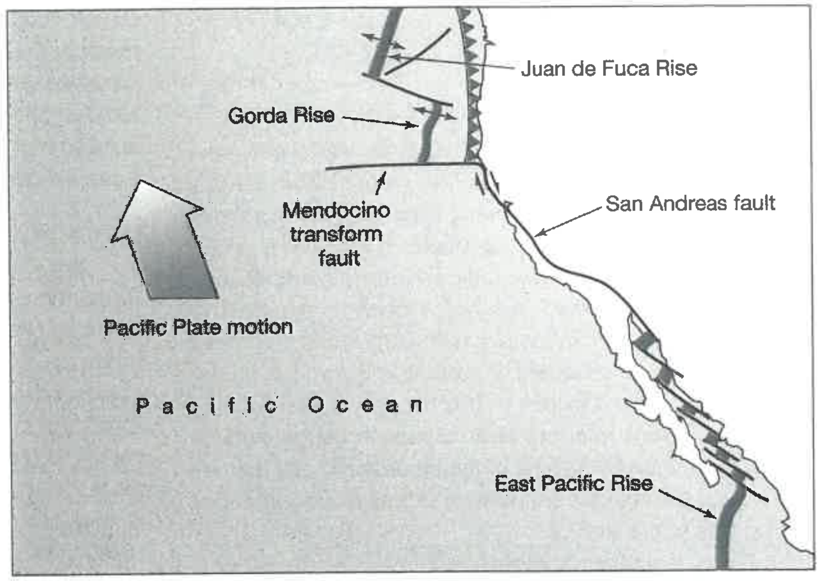


FIGURE 7-20 Transform faults. Seafloor generated at the Juan de Fuca Ridge moves southeastward, past the Pacific plate and beneath the North American plate, at the Mendocino transform fault. This fault connects a divergent boundary to a subduction zone. The San Andreas Fault, another transform fault, forms the connection between two spreading centers: the Juan de Fuca Ridge and a divergent zone in the Gulf of California. (*Source:* From J. P. Davidson, W. E. Reed, and P. M. Davis, *Exploring Earth: An Introduction to Physical Geology*, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

As Figure 7-21 shows, some lithospheric plates consist of oceanic lithosphere welded to continental lithosphere. The ocean–continent boundary is not a plate boundary at all. The North American plate, for example, is made up of the North American continent and oceanic lithosphere to the east. The North American plate continues beyond the continent–ocean lithospheric boundary to the axis of the Mid-Atlantic Ridge (the mid-ocean ridge down the center of the Atlantic Ocean). In such situations, the ocean widens as the continents drift away from the mid-ocean ridge; the continental margin is referred to as passive. *Passive continental margins* consist of a broad, gently seaward-dipping continental shelf that gives way to the more steeply dipping continental slope. The continental lithosphere is welded to the oceanic lithosphere beneath the continental shelf. In other cases, such as along much of the Pacific Ocean, the ocean–continent boundary is a convergent margin and is therefore a site of subduction; such margins are said to be active. On *active continental margins,* including that off the Pacific coast of the northwestern United States, the continental shelf is narrow.

The oceanic ridge is offset in numerous places by transform faults. The relative motion across these faults is parallel to the plate boundary. Transform faults generate considerable amounts of earthquake activity. Around the globe, convergent and divergent margins are connected by transform faults.

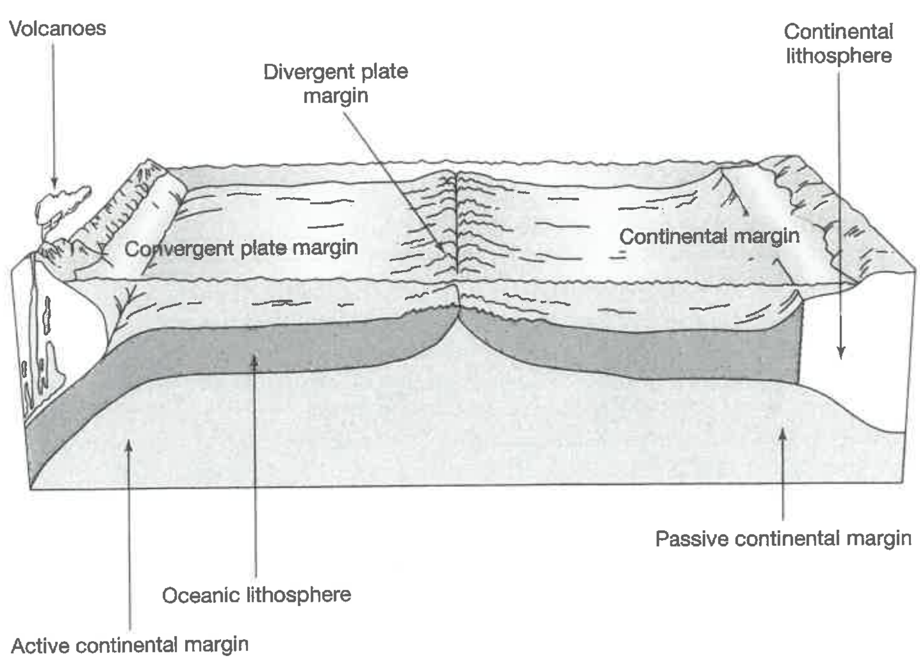


FIGURE 7-21 Schematic view of the relationships among the types of plate boundaries. (*Source:* From J. P. Davidson, W. E. Reed, and P. M. Davis, *Exploring Earth: An Introduction to Physical Geology*, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

THE PHYSIOLOGY OF THE SOLID EARTH: WHAT DRIVES PLATE TECTONICS?

Heat from the Deep

In the simplest sense, plate tectonics is the surface expression of the mechanism by which heat escapes from Earth's interior. Although there are spatial variations, temperatures generally increase through the mantle. This heat is transported to the surface, where it escapes to the atmosphere. (The average *geothermal heat flux,* or heat transported to the surface, is 0.06 W/m^2, which is trivially small compared to the net absorbed solar flux of about 240 W/m^2, the energy budget discussed in Chapter 3. It is the only heat available, though, in Earth's interior.) Heat is transported by convection in the mantle to the base of the lithosphere, and then by conduction through the lithosphere or convection at midocean ridges to the surface.

What is the origin of this heat in Earth's interior? It comes from two major sources: (1) radioactive decay and (2) residual heat from Earth's formation. A third source, the growth of the inner core (discussed previously), drives convection of the outer core but is only a small contributor to the energy budget of Earth's interior.

RADIOACTIVE DECAY We discussed the fundamentals of radioactive decay back in Chapter 5. There, we applied these concepts to the decay of carbon-14. The important radioactive elements in the solid Earth are potassium, uranium, and thorium. Their *half-lives* are on the order of hundreds of millions to billions of years (whereas carbon-14, if you recall, has a half-life of only 5,730 years; see Chapter 5). Thus, the decay rates of these isotopes are quite low. However, the crust and mantle contain significant concentrations of these elements, so their radioactive decay generates a considerable amount of heat. Because radioactive decay leads to a continuous loss of radioactive materials from Earth's interior, the abundance of these materials must have been greater in the past than it is now. Similarly, the rate of heat production (and of heat loss) must have been much higher in the past. On the basis of the abundance of potassium, uranium, and thorium in the crust and mantle, we can calculate that the amount of radioactive heat production has decreased by about a factor of 5 since Earth formed 4.6 billion years ago.

OTHER HEAT SOURCES Other sources of heat are residual; they are associated with heating events during Earth's formation. As we mentioned briefly earlier, Earth (and the other planets) was formed by the accretion of larger and larger clumps of matter into moon-sized objects called *planetesimals.* The planetesimals collided and merged, forming a large, primitive planet. A tremendous amount of energy was transferred to Earth during the accretion of the planet by collisions with planetesimals. The larger collisions probably caused widespread melting of Earth's upper mantle. The segregation of Earth into a less-dense mantle and crust and a denser core released gravitational energy in the form of heat. Convection of the outer core and mantle has been transferring this heat to Earth's surface ever since.

Convection in the Mantle

But how can a solid convect? Convection generally is thought of as a process that affects fluids. Yet solids need not be rigid; witness the flow of glaciers or the ductility of plastics. Rocks are ductile at the temperatures and pressures that occur within the mantle. When heated locally, these materials expand, become less dense, and rise buoyantly, although very slowly. Cooler, denser material sinks and replaces the buoyant material. In this way, mantle rocks can convect. Upon rising to the base of the lithosphere, the material cools as heat is transferred conductively to the lithosphere. As the material continues to cool, it travels laterally. It cools so much that it eventually becomes denser than the underlying lithosphere and descends. Thus, the material sinks back into the asthenosphere. The cycle continues as the material is again heated and becomes buoyant.

A CLOSER LOOK

Radiometric Age Dating of Geological Materials

Suppose that we have a rock sample that contains both a radioisotope and its decay product, and we know exactly how much of that isotope existed when the material formed. If we also know the half-life of that isotope, then we can calculate the age of the sample. This method, called *radiometric age dating*, has proved extremely useful in providing absolute dates for the geological time scale and for other specific events in Earth's history. The accuracy of radiometric dating drops rapidly after eight or nine half-lives. For example, the radioactive isotope of carbon, ^{14}C or *radiocarbon*, has a half-life of just 5,730 years. Thus, *radiocarbon dating* is accurately applied only to samples less than a few tens of thousands of years old. The long-lived radioisotopes of potassium and uranium, however, are useful for determining the ages of the oldest rocks on Earth, of lunar material, and of meteorites. Especially useful are the radioactive isotopes of uranium, ^{238}U (half-life of 4.5 billion years) and ^{235}U (half-life of 0.713 billion years), which decay through a series of intermediate steps to stable lead isotopes ^{206}Pb and ^{207}Pb, respectively. The radioactivity of the uranium in these ancient rocks is not high enough to be measured directly, but we can use an instrument called a *mass spectrometer* to determine the relative amounts of the lead isotopes. From these ratios and the known half-lives, we can accurately date the rocks.

The size of mantle convection cells is unknown. Smaller cells may be generated separately within the upper mantle and within the lower mantle, or the whole mantle below the lithosphere may be involved. The nature of the mantle transition zone is the distinguishing factor between the two convection mechanisms. If this zone marks a change in chemical composition, then convection cells do not cross it because if they did, the compositional distinctions would be lost. In this case, there are likely to be separate convective cells in the upper and lower mantle (Figure 7-22a). Conversely, if the transition zone is the result of mineralogical rather than chemical changes, and if these changes take place quickly relative to the rate of convection, whole-mantle convection is possible (Figure 7-22b).

The lithosphere is an integral part of the mantle convection system. In a sense, the lithosphere is the cool upper boundary of the convective cell. However, the subduction of cool oceanic lithosphere undoubtedly perturbs the internal thermal structure of the upper mantle and the distribution of convective cells. Oceanic lithosphere is so dense that a slab of oceanic lithosphere at a subduction zone may sink to great depths within the mantle, become detached from the surface portion of the plate, and actually cool regions of the mantle from below. Such regions would become thermally stable, preventing convection locally. Lateral movements of the plates might also induce lateral movements of the underlying asthenosphere. And finally, the separation of the lithosphere at seafloor spreading centers might drive the mantle to upwell. As the asthenosphere rises, it expands and melts, which further enhances the upwelling of more magma. Thus there are a number of lithosphere–asthenosphere interactions that affect the nature of convection in the mantle.

Forces Acting on Plates

The most important lithosphere–asthenosphere interaction is that which drives the motion of the plates. When plate tectonics was first proposed, *mantle drag*, or friction between the convecting asthenosphere and the overlying rigid lithosphere, was considered to be the cause of plate motions (Figure 7-23). Now geologists recognize a number of other forces that act on plates. These forces include

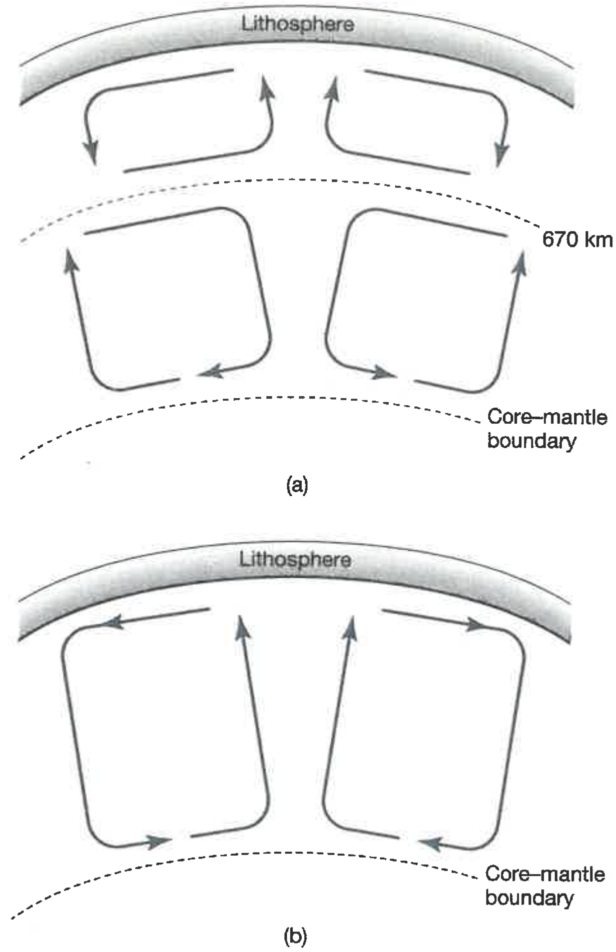


FIGURE 7-22 Mantle convection may (a) separate into upper and lower mantle convective cells or (b) involve the whole mantle. (*Source:* From S. Judson and S. M. Richardson, *Earth: An Introduction to Geologic Change*, 1995. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

the gravitational "push" generated by the high topography of a mid-ocean ridge on the rest of the oceanic plate (ridge push); the increasing density of the oceanic lithosphere as it cools, which pulls the opposite end of the plate into a subduction zone (slab pull); the elastic resistance of the oceanic plate to being bent into a subduction zone

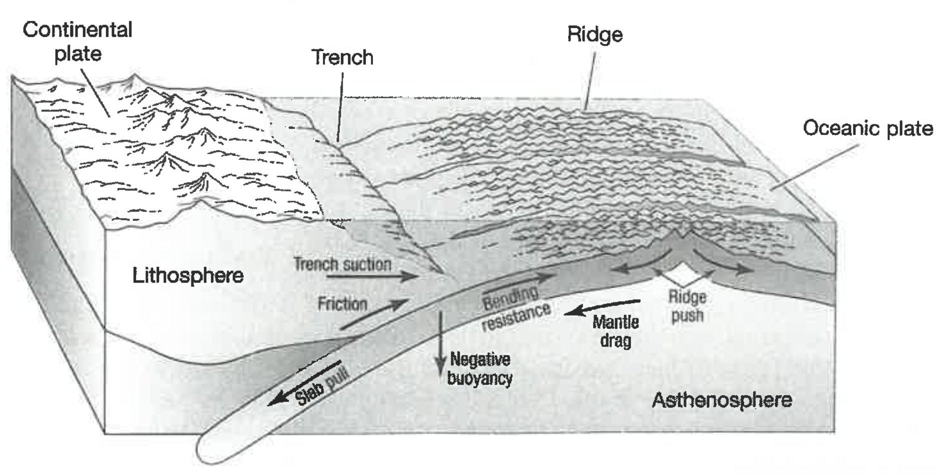


FIGURE 7-23 The various forces acting on plates at their leading and trailing edges. The motion of the plates responds to the sum of these forces. See the text for a discussion of the origin of each force. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems*, 8/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

(bending resistance); the tendency for the overriding plate to be drawn toward a subduction zone as the subducting slab bends (which otherwise would move the trench away from the overriding plate (trench suction); friction between the subducting slab and the overlying lithosphere (friction); and a tendency for the oceanic plate to sink as it cools and becomes denser (negative buoyancy). The overall motion of a given plate is the result of the balance of all these forces. Analysis of plate motions today argues for a predominant role played by the push of the ridges and the pull of the subduction zone, but in the past, mantle drag may have played a more important role in the rifting apart of supercontinents (see "Evolution of the Driving Force").

RECYCLING OF THE LITHOSPHERE: THE ROCK CYCLE

All rocks of the lithosphere ultimately derive from igneous rocks. Igneous rocks of the oceanic lithosphere are born when extruded volcanically at mid-ocean ridges, and they die on average some 80 million years later when subducted and incorporated into the asthenosphere. (The oldest oceanic crust is about 200 million years old.) In contrast, the oldest-known continental rocks formed nearly 4 billion years ago. These record-setters reside in the old, previously active but now tectonically dormant regions of the continental interiors known as **cratons**. These very old continental blocks form the nucleus on which subsequent plate collisions have plastered on new material over the past 3–4 billion years, leading to the growth of the continents (Figure 7-24).

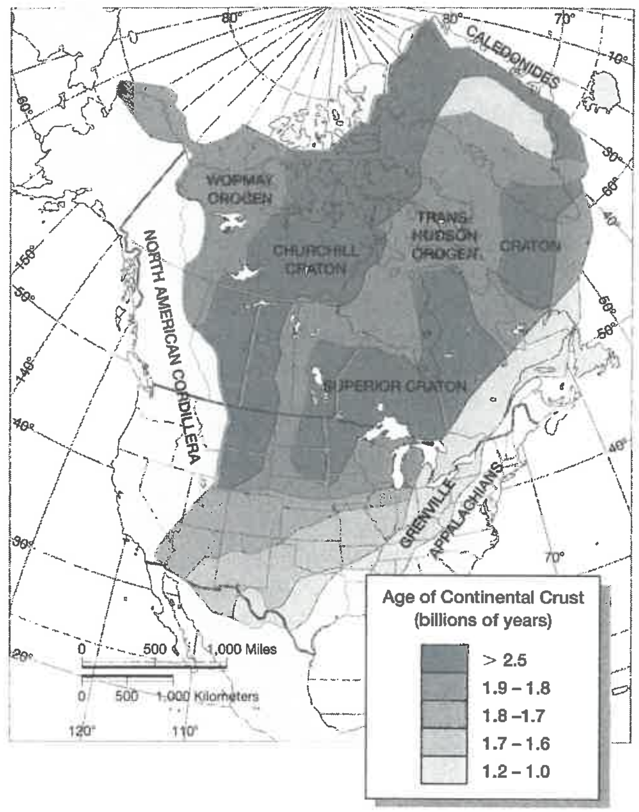


FIGURE 7-24 The ages of the components of the North American continent reveal that the continent has grown by the amalgamation of very old cratons, followed by accretion of younger material onto the periphery of the craton during plate collisions. (*Source:* From J. P. Davidson, W. E. Reed, and P. M. Davis, *Exploring Earth: An Introduction to Physical Geology*, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

Weathering and Erosion

Once formed, igneous rocks are subject to a variety of processes that can alter their chemical composition and weaken their structural integrity, typically leading to complete disintegration or *dissolution* (dissolving away) of the original rock. On the seafloor, oceanic lithosphere is altered as hydrothermal solutions circulate the oceanic crust, as previously discussed, but these processes tend not to destroy the rocks. In contrast, igneous rocks that are exposed on land are subject to a variety of physical, biological, and chemical forces that tend to degrade the rock. These are referred to as *weathering* processes. Weathering transforms solid rock into small particles (sediments) and dissolved material.

A number of processes contribute to physical weathering. Rocks expand and fracture as the weight of overlying material is removed through erosion. In temperate latitudes, water seeps into these fractures in spring, summer, and fall. It then expands when it freezes during the winter, cracking the rock. Rocks exposed at the surface in high latitudes or at high altitudes are ground up as glaciers advance and retreat. Finally, there are biophysical mechanisms of weathering, including the action of plant roots that penetrate along these fractures, wedging the rock apart.

Chemical weathering results from the tendency for minerals to dissolve when exposed to rainwater and acidic soil waters generated by bacteria, fungi, or plant root discharges. The products of these chemical reactions include dissolved materials and relatively insoluble clays that form in the soils.

The transport of the products of weathering to basins where sediment accumulates is called **erosion**. In this process, crustal materials, decomposed and loosened by weathering, and the clays that form in the soils are transported by winds, landslides, and streams to sites of deposition. These sites include lakes and flood plains on land and deltas and deeper basins in the ocean. In the process, landscapes are created: Valleys form as erodable material is removed, leaving peaks and ridges of material that are resistant to erosive forces.

Sediment Accumulation

The accumulation of sediments depends on two factors: the rate of supply of sediment and the amount of space available to accumulate the sediment. The great depth of water in deep marine basins allows for the thick accumulation of sediments. As the sediments accumulate, however, the seafloor rises toward sea level, and the amount of available space diminishes. Sedimentary deposits also form in shallow-water settings, though, such as the margins of the Gulf of Mexico, where subsidence of the seafloor (resulting from tectonic forces that stretch and thin the lithosphere, or cooling and contraction) allows for continued accumulation of sediment.

As sediments continue to accumulate in these basins, the material is compacted by the increasing weight of overlying sediments. That weight can become so great that fluids trapped between the sediment grains are expelled. Eventually, as the burial process continues, sediments may become buried to a depth of several kilometers below the seafloor. At these depths, temperatures can exceed 200°C, and the pressures can be hundreds of times the atmospheric pressure. In addition, the fluids that circulate through these buried sediments are quite distinct chemically from surface waters. As a result of these environmental changes, sediments undergo further compaction, and the small voids that remain between sediment grains become filled with mineral cements precipitated from the subsurface fluids. In other words, the sediments lithify and become sedimentary rocks.

Uplift

If the continents were at sea level, there would be virtually no driving force for weathering and erosion, and thus no rock cycle. However, continental collisions crumple the crust, producing mountains and high plateaus, and oceanic collisions produce volcanic islands. These plate collisions generate the topography that is then destroyed by erosion. The elevation of a mountain range represents the competition between uplift and erosion. Young mountain belts like Taiwan or Papua New Guinea are undergoing rapid uplift. Steep mountain slopes develop that stimulate rapid erosion as well, but uplift exceeds erosion, and elevations grow. Old mountain belts such as the Appalachians of the eastern United States are not being rapidly uplifted. Erosion is dominating, and the high peaks of the geologic past are gone, having eroded to low ridges.

Erosion promotes the recycling of sedimentary rocks before metamorphism or melting occurs. Sedimentary deposits situated along active continental margins can be entrained into the convergent plate motions and uplifted onto the continents. As soon as the sedimentary rocks become exposed at the surface, weathering and erosion commence.

Metamorphism and Melting

If instead sedimentary (or igneous) rocks are subjected to high temperatures and pressures in Earth's interior, they metamorphose. If they ultimately melt, they form magma that can itself ascend and become igneous rock. Metamorphism occurs deep in sedimentary basins along passive margins, in the deformed regions of active margins, and in other regions of the crust where igneous activity has injected hot igneous rock into what was cooler crustal rock (igneous or sedimentary). The generation of magma from crustal rocks typically occurs only where those rocks have been carried deep within Earth's interior, for example, in subduction zones.

The Rock Cycle

There are a number of alternative pathways in this process, but overall the complete regeneration of rock is called the **rock cycle** (Figure 7-25). The rock cycle is a consequence of plate tectonics. One complete cycle takes about 100 million years. However, the average lifetime of continental lithosphere as a whole is actually much longer (a few hundred million years), because the interiors of the continents are well insulated from the tectonic activity that occurs along their margins.

It is important to recognize that the rock cycle is not completely closed: New crustal material is produced through the emplacement of magmas derived from the mantle, and older crustal materials are taken back to the mantle at subduction zones. This slow exchange between the mantle and crust replenishes the crust, on average,

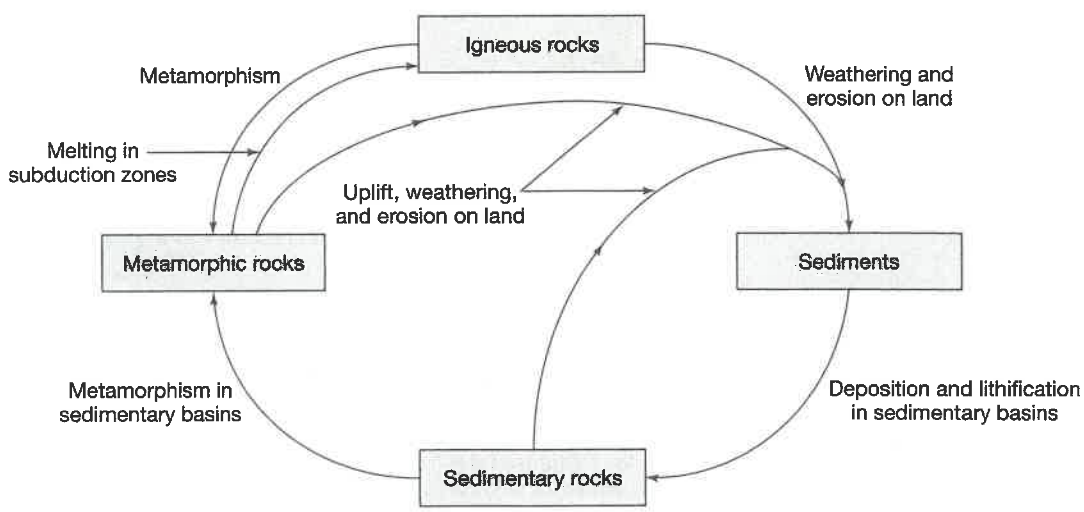


FIGURE 7-25 The rock cycle.

every 2–3 billion years. In other words, a large proportion of the geologic record of the early Earth's crust has not only been through the rock cycle many times, but has been ingested (perhaps once and for all) into Earth's interior. Nevertheless, averages can be deceptive: The interior parts of the continents (the cratons) can be billions of years old and represent crustal material that has never been recycled, whereas the continental margins are recycled on a time scale much shorter than a billion years.

PLATE TECTONICS THROUGH EARTH HISTORY

Evolution of the Driving Force

Earth has been losing heat throughout its 4.6 billion-year history. Although early in this history other mechanisms may have dominated, for at least the past 4 billion years heat loss has occurred by mantle convection. The rate of heat loss on the early Earth was several times the present value, presumably fueling higher relative rates of seafloor production and subduction. But the style of subduction, and the balance of forces acting on the plates, may have been somewhat different on the early Earth than they are today. Presuming that plate velocities of a few centimeters per year have prevailed over the past 4 billion years or so, the continents have moved great distances during that time.

Wilson Cycles

A pattern seems to be emerging: Continents assemble into a supercontinent, which then breaks apart; these smaller continents eventually disperse and then reassemble. This plate tectonic cycle has been dubbed the **Wilson cycle**, in honor of one of the pioneers of plate tectonics, Canadian geologist J. Tuzo Wilson. From paleogeographic reconstructions it appears that the cycle of supercontinent assembly and destruction takes about 500 million years. We can also approximate the duration of a Wilson cycle from the time it takes a plate to make its way halfway around Earth. Taking a typical plate speed of 4 cm/yr (or 40 km/million yr) and a half-circumference (at the equator) of around 20,000 km, we confirm that two continents rifting apart would meet each other again, on the far side of the planet, in around 500 million years. Accordingly, given that Pangea formed about 300 million years ago, the next supercontinent should be formed in about 200 million years as the Pacific Ocean closes, swallowed up by the subduction zones that surround it.

Why do all the continents come together into a supercontinent rather than displaying a less organized, more random pattern of collision and rifting apart? Of course, drifting continents on a finite globe are bound to collide, so larger continents are likely to form. A somewhat controversial hypothesis argues instead that the continents are drawn toward cold regions of the asthenosphere (Figure 7-26). Once assembled, the thick supercontinent acts as an insulator, slowing the release of heat from the mantle. Mantle temperatures rise beneath the supercontinent, modifying the pattern of convection. The resulting tension at the surface eventually rips the supercontinent apart. The continents begin to move from this region of hot upwelling mantle to a site thousands of kilometers away, where the mantle has cooled and downwelling has commenced. This hypothesis is consistent with present-day plate speeds and seismic tomography, which together suggest that the continents, with the exception of Africa, appear to be moving toward regions of cold mantle. Africa is situated above hot mantle, as evidenced by the East African Rift zone, a place where tension within the continental lithosphere is creating a rift in the continent. Africa has apparently moved little since the breakup of Pangea, and the underlying mantle still retains the heat built up during Pangea's existence.

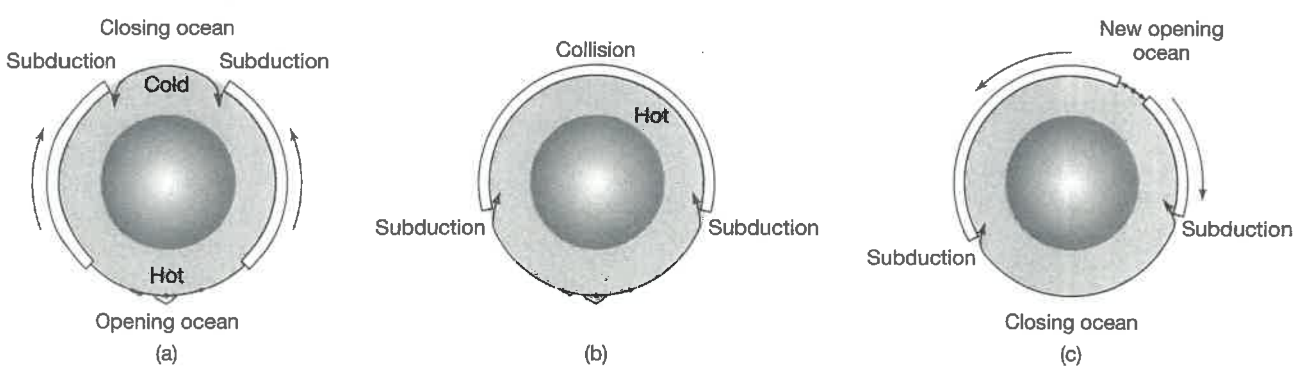


FIGURE 7-26 The Wilson cycle of supercontinent assembly and fragmentation. (a) The continents are drifting toward a region of cold asthenosphere. The closing ocean is lined by subduction zones and is contracting. The other ocean is opening, and the oceanic lithosphere is connected to the continental lithosphere at both margins. (b) The continental fragments have collided, forming a supercontinent. Subduction has begun along the margins of the formerly opening ocean. The insulating effects of the thick continental lithosphere lead to the buildup of heat and the initiation of rifting. (c) What once was an opening ocean has become a closing ocean, with cool asthenosphere beneath. One Wilson cycle is now complete. (*Source:* P. Kearey and F. J. Vine, *Global Tectonics*, Oxford: Blackwell Scientific, 1990.)

Chapter Summary

1. The solid Earth is dynamic, not static. Wegener's idea of drifting continents proposed in the early 20th century has largely been substantiated.
   a. New seafloor is created at mid-ocean ridges and moves outward as ocean basins grow.
   b. Old seafloor is destroyed at deep-sea trenches in subduction zones. Earthquakes outline the surface of the slab of oceanic lithosphere being subducted beneath the continent; their foci along a continental margin become ever deeper away from the trench.
   c. Deeper probing of Earth's interior has revealed heterogeneity in composition and temperature that can be the result only of large-scale circulation in the mantle and outer core.
      i. This circulation is fueled by residual heat from the formation of the planet 4.6 billion years ago and by heat that continues to be produced as the result of the radioactive decay of potassium, uranium, and thorium in the mantle and crust.
      ii. Mantle circulation is the result of convection, not unlike that of the troposphere. Scientists continue to debate whether convective cells extend throughout the mantle or whether a dual system of upper and lower mantle convection is in operation.
2. The signature of plate tectonics is best seen at the surface, along the margins of lithospheric plates.
   a. At plate margins, divergent, convergent, and transform plate motions generate impressive topographic features: mid-ocean ridges and continental mountain belts and volcanoes, deep-sea trenches, and transform faults, respectively. These features themselves change with time.
      i. Mountains grow through plate collision but shrink through weathering and erosion.
      ii. Sediments are transported to the oceans, where they fill in deep-sea trenches as well as basins generated through subsidence.
      iii. Burial converts sediments to sedimentary rocks, but in time these rocks are likely to become reexposed as the result of plate convergence and uplift, or metamorphosed and perhaps ultimately melted.
   b. This cycle of rock production and destruction (the rock cycle), driven largely by plate tectonic processes, continuously resurfaces the planet and recycles material between the crust and mantle.
3. The movement of lithospheric plates is driven by forces at plate margins and, at the base of the lithosphere, by friction with the convecting asthenosphere.
   a. Magnetic stripes displaying mirror-image patterns across the mid-ocean ridges document the creation of seafloor at the ridges.
   b. Plate motions are slow on human time scales (on the order of centimeters per year), but over geological time they can lead to the complete redistribution of continents on the globe.
      i. These movements appear to be organized, following a pattern called Wilson cycles. These cycles consist of the alternating assembly of supercontinents (perhaps at the position of mantle downwelling) and their subsequent breakup (as the insulating effects of the supercontinent lead to sublithospheric heating and mantle upwelling).
      ii. The most recent supercontinent, Pangea, formed more than 300 million years ago. Its breakup, beginning some 200 million years ago, led to the creation of the Atlantic Ocean and the separation of North America and South America from Europe and Africa.
      iii. The Pacific Ocean is currently shrinking, and it is estimated that in another 200 million years or so the continents will again reassemble into another supercontinent.

Key Terms

asthenosphere
basalt
body waves
chemosynthesis
continental drift
craton
earthquake
erosion
granite
igneous rock
lithification
lithosphere
magma
magnetic dynamo
metamorphic rocks
mid-ocean ridge
mineral
Moho
Pangea
plate tectonics
polarity
P wave
rock cycle
seafloor spreading
sedimentary rock
sediments
seismic wave
silicate mineral
slab
subduction
surface waves
S wave
transform fault
Wilson cycle

Review Questions

1. Why was the theory of continental drift not immediately embraced by the scientific community in the 1920s?
2. What is the Moho?
3. What are the bases for the two major divisions of Earth's interior—one that distinguishes crust, mantle, and core and the other that distinguishes lithosphere and asthenosphere?
4. Compare and contrast P and S seismic waves.
5. Why are earthquakes focused along plate margins?
6. What are the sources of heat in Earth's interior?
7. What is magnetic polarity? What role did it play in the generation of ideas regarding seafloor spreading?
8. What are the three types of plate boundaries, and what surface features are characteristic of each?
9. What is erosion?
10. How can radioactivity be used to determine the age of a rock?
11. What are the driving forces for plate movement?
12. What is hypothesized to drive the Wilson cycle of plate fragmentation and reassembly?

Critical-Thinking Problems

1. We have seen that cooling of the oceanic lithosphere causes contraction, leading to subsidence of the seafloor away from the axis of spreading. The depth d of the ocean floor, measured in meters, increases with age t, measured in millions of years from the present, according to the following equation (valid for seafloor younger than 80 million years old):

$$d = 2500 + 350 \cdot \sqrt{t}$$

Graph a cross section of a mid-ocean ridge that is spreading symmetrically in both directions at a rate of 1 cm/yr (10 km/million yr). The age of the oldest seafloor shown should be 80 million years.

2. Duplicate Figure 7-10 and answer the following questions:
   a. Draw a line from the tip of Florida horizontally across the Atlantic to northwest Africa, a distance of about 6400 km. Now, graph the age of the seafloor (on the y-axis) against the distance from the ridge axis (on the x-axis). From this graph, determine the spreading rate for each geologic interval represented, averaging the two values determined for eastward and westward spreading. Graph these values (y-axis) as a function of time (in million years, on the x-axis).
   b. How has the Atlantic spreading rate varied over the last 200 million years?

Further Reading

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