PART D

TECTONICS

CHAPTER FOURTEEN

Whole-Earth Structure and Plate Tectonics

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14.1 INTRODUCTION

Up to this point in the book, we have focused on understanding the nature and origin of specific structures (such as faults, folds, and fabrics), and on characterizing relations among deformation, stress, and strain. This discussion has emphasized features that we see at the microscopic to outcrop scale. Now, let's change our focus to the *regional* scale by discussing tectonics.[1] Geologists use the term **tectonics,** in a general sense, to refer to the sum of physical processes that yield regional-scale geologic features (Figure 14.1). Studies in tectonics consider such issues as the origin of mountain belts, the growth of continents, the formation of the ocean floor, the development of sedimentary basins, and the causes of earthquakes and volcanoes. In the next six chapters (Part D), we introduce these "big-picture" issues, and show how rock deformation occurs hand-in-hand with many other geologic phenomena as the Earth evolves.

This chapter covers two topics that provide the foundation for Part D of this book. We begin by describing **whole Earth structure,** meaning, the internal layering of the Earth. Geologists subdivide the Earth's insides into layers in two different ways. First, based on studying the velocity of seismic waves that pass through the Earth, we subdivide the *entire* planet into the **crust,** the **mantle,** and the **core,** named in sequence from surface to center. Second, based on studying rock rheology (the response of rock to stress), we subdivide the *outer several hundred kilometers* of the Earth into the lithosphere and asthenosphere. The **lithosphere** is the outermost of these **rheologic layers** and forms Earth's rigid shell. The lithosphere, which

[1]The word "tectonics" was derived from the Greek *tektos,* meaning "builder."

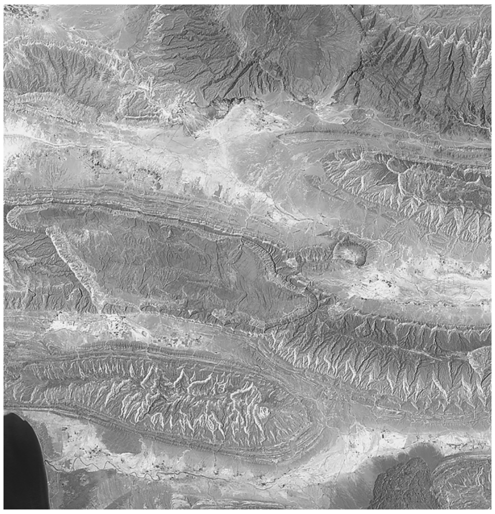


FIGURE 14.1 Landsat image of tectonically active region, the Zagros Mountains of Iran. The dome shapes are relatively recent developed anticlines. The topography clearly reflects the underlying structure.

consists of the crust *and* the outermost mantle, overlies the **asthenosphere,** a layer that behaves plastically (meaning that, though solid, it can flow). The asthenosphere lies entirely within the mantle. A background on whole Earth structure sets the stage for introducing **plate tectonics theory** (or simply, "**plate tectonics**"). According to this theory, the lithosphere consists of about 20 discrete pieces, or **plates,** which slowly move relative to one another. In our discussion, we describe the nature of plates, the boundaries between plates, the geometry of plate movement, and the forces that drive plate movement.

14.2 STUDYING EARTH'S INTERNAL LAYERING

Before the late nineteenth century, little was known of the Earth's interior except that it must be hot enough, locally, to generate volcanic lava. This lack of knowledge was exemplified by the 1864 novel *Journey to the Center of the Earth,* in which the French author Jules Verne speculated that the Earth's interior contained a network of caverns and passageways through which intrepid explorers could gain access to the planet's very center. Our picture of Earth's insides changed in the late nineteenth century, when researchers compared the gravitational pull of a mountain to the gravitational pull of the whole Earth and calculated that our planet has a mean density about 5.5 g/cm^3, more than twice the density of surface rocks like granite or sandstone. This fact meant that material inside the Earth must be much denser than surface rocks—Verne's image of a Swiss-cheese-like Earth could not be correct.

Once researchers realized that the interior of the Earth is denser than its surface rocks, they worked to determine how mass is distributed inside the Earth. First, they assumed that the increase in density occurred gradually, due entirely to an increase in **lithostatic pressure** (the pressure caused by the weight of overlying rock) with depth, for such pressure would squeeze rock together. But calculations showed that if density increased only gradually, so much mass would lie in the outer portions of the Earth that centrifugal force resulting from Earth's spin would cause our planet to flatten into a disc. Obviously, such extreme flattening hasn't happened, so the Earth's mass must be concentrated toward the planet's center. This realization led to the image of a **layered Earth,** with a very dense central region called the "core," surrounded by a thick "mantle" of intermediate density. The mantle, in turn, is surrounded by a very thin skin, the relatively low density "crust." Eventually, studies showed that the density of the core reaches 13 g/cm^3, while crustal rocks have an average density of 3–6 g/cm^3.

In the twentieth century, geoscientists have utilized a vast array of tools to provide further insight to the mystery of what's inside the Earth. Our modern image of the interior comes from a great variety of data sources, some of which are listed in Table 14.1. Standard textbooks on geophysics (see Additional Reading) will explain these in greater detail. Here, we focus on constraints on Earth's structure provided by seismic data.

14.3 SEISMICALLY DEFINED LAYERS OF THE EARTH

Work by seismologists (geoscientists who study earthquakes) in the first few decades of the twentieth century greatly refined our image of the Earth's interior. **Seismic waves,** the vibrations generated during an earthquake, travel through the Earth at velocities ranging from about 4 km/s to 13 km/s. The speed of the waves, their **seismic velocity,** depends on properties (e.g., density, compressibility, response to shearing) of the material through which the waves are traveling. When waves pass from one material to another, their velocity changes abruptly, and the path of the waves bends. You can see this phenomenon by shining a flashlight beam into a pool of water; the beam bends when it crosses the interface between the two materials.

TABLE 14.1 SOURCES OF DATA ABOUT THE EARTH'S LAYERS

Drilling	Deep holes have been drilled in the continents and in the ocean floor (Figure 14.2a). Samples extracted from these holes, and measurements taken in these holes, provide constraints on the structure of the uppermost part of the Earth. But even the deepest hole on Earth, drilled in Russia, penetrates only about 15 km deep, a mere 0.2% of the Earth's radius.
Electrical conductivity	The manner in which electricity conducts through subsurface Earth materials depends on rock composition and on the presence of fluids (water and oil near the surface, magma at depths). Electrical conductivity measurements can detect the presence of partially molten rock (rock that is starting to melt, so that it consists of solid grains surrounded by films of melt) in the mantle and crust.
Earth's density	By measuring the ratio between the gravitational force generated by a mountain (of known dimensions and composition) and the gravitational force generated by the whole Earth, geoscientists determined the mass of the Earth and, therefore, its average density. Knowledge of density limits the range of possible materials that could comprise the Earth.
Earth's shape	The Earth is a spinning sphere. Thus, Earth must be largely solid inside, for if its insides were liquid, its spin would cause the planet to be disc-shaped. Similarly, if density were uniform through the planet, Earth's spin would cause the planet to be more flattened than it actually is. Thus, the shape of the Earth requires that it have a core that is denser than the surrounding mantle and crust.
Exposed deep crust	In ancient mountain belts, the combined process of faulting, folding, and **exhumation** (removal of overlying rock) have exposed very deep crustal levels. In fact, outcrops in the interior of large mountain ranges contain rocks that were once at depths of 20 to 50 km in the crust. In a few localities geologists have even found rocks brought up from about 100 km depth (these rocks contain an ultrahigh-pressure mineral named coesite). Study of these localities gives a direct image of the geology at depth.
Geochemistry	Rocks exposed at the surface were, ultimately, derived by extraction of melt from the mantle. Thus, study of the abundance of elements in rocks at the surface helps to define the range of possible compositions of rocks that served as the source of magma at depth.
Gravity field	Measurements of variations in the strength of Earth's gravity field at the surface give a clue to the distribution of rocks of different density below the surface, for denser rocks have more mass and thus cause greater gravitational attraction. Gravity measurements can indicate where the lithosphere is isostatically compensated and where it is not (see our discussion of isostasy later in this chapter for a further explanation).
Lab experiments	Lab studies that determine the velocity of seismic waves as a function of rock type, under various conditions of pressure and temperature, allow geoscientists to interpret the velocity versus depth profile of the Earth, and to interpret seismic-refraction profiles. Sophisticated studies of elastic properties of minerals squeezed in diamond anvils and heated by lasers even allow the study of materials at conditions found in the lower mantle or core.
Lithospheric flexure	The lithosphere, the outer relatively rigid shell of the Earth, bends ("flexes") in response to the addition or removal of a surface load. For example, when a huge glacier spreads out over the surface of a continent during an ice age, the surface of the continent bends down, and when the glacier melts away, the surface slowly rises or "rebounds." When the lithosphere bends down, the underlying asthenosphere must flow out of the way, and when rebound occurs, the underlying asthenosphere flows back in. Thus, the rate of sinking or rebound depends on the rate at which the asthenosphere moves, and thus on the viscosity (resistance to flow) of the asthenosphere. As another example, the shape of a lithosphere where it bends down into the mantle at convergent plate boundaries (subduction zone) provides insight into flexural strength ("bendability") of the plate. Thus, studies of flexure give insight into the rheology of crust and mantle (e.g., they tell us if it is it elastic, viscous, or viscoelastic).
Magnetic anomalies	We say that a magnetic anomaly occurs where the measured strength of the Earth's magnetic field is greater or lesser than the strength that would occur if the field were entirely due to the Earth's internal field (caused by the flow of iron alloy in the outer core). Anomalies occur because of the composition of rock in the crust, or due to the polarity of the magnetic field produced by tiny grains of iron-bearing minerals in a rock.

TABLE 14.1 SOURCES OF DATA ABOUT THE EARTH'S LAYERS

Meteorites	Meteorites are chunks of rock or metal that came from space and landed on Earth. Some are relict fragments of the material from which planets first formed, while others are fragments of small planets that collided and broke apart early in the history of the solar system. Thus, some meteorites may be samples of material just like that which occurs inside the Earth today.
Ophiolites	An ophiolite is a slice of oceanic crust that was thrust over continental crust during collisional orogeny, and thus is now exposed on dry land for direct examination by geologists. Study of ophiolites gives us an image of the structure of the oceanic crust.
Seismic reflection	Geoscientists have developed methods for sending artificial seismic waves (vibrations generated by explosions or by large vibrating trucks) down into the crust and upper mantle. These waves reflect off boundaries between layers in the subsurface and then return to the surface. Sophisticated equipment measures the time it takes for this process to occur and, from computer analysis of this data, geoscientists can create cross sections of the subsurface that reveal formation contacts, folds, faults, and even the Moho.
Seismic refraction	When a seismic wave reaches the boundary between two layers, some of the energy reflects, or bounces off the boundary, while some refracts, meaning that it bends as it crosses the boundary. Studies of refracted waves can be used to define the velocity of seismic waves in a layer. Such studies provide insight into the composition and dimensions of subsurface layers.
Seismic-wave paths	By studying the paths that earthquake waves follow as they pass through the Earth, and the time it takes for the waves to traverse a distance, geoscientists can identify subsurface layer boundaries and layer characteristics.
Seismic tomography	New computer techniques, similar to those used when making medical "CAT scans," allow geoscientists to create a three-dimensional picture of seismic velocity as a function of location in the crust, mantle, and inner core (see Figure 14.9). These images can be interpreted in terms of variation in material properties controlled by temperature and/or chemistry.
Xenoliths	Xenoliths (from the Greek "xeno-," meaning foreign or strange) are preexisting rocks that have been incorporated in a magma and brought to or near the Earth's surface when the magma flows upward. Some xenoliths are fragments of the deep crust and/or upper mantle, and thus provide samples of these regions for direct study (Figure 14.2b).

(a) (b)

FIGURE 14.2 (a) Photograph of the *Joides Resolution,* a drilling ship used to drill holes in the sea floor for research purposes. The ship has drilled hundreds of holes all over the Earth, allowing geologists to understand the layering of the ocean crust. (b) Photo of a block of basalt containing xenoliths of dunite. Dunite, a variety of mantle peridotite, consists almost entirely of small, green olivine crystals.

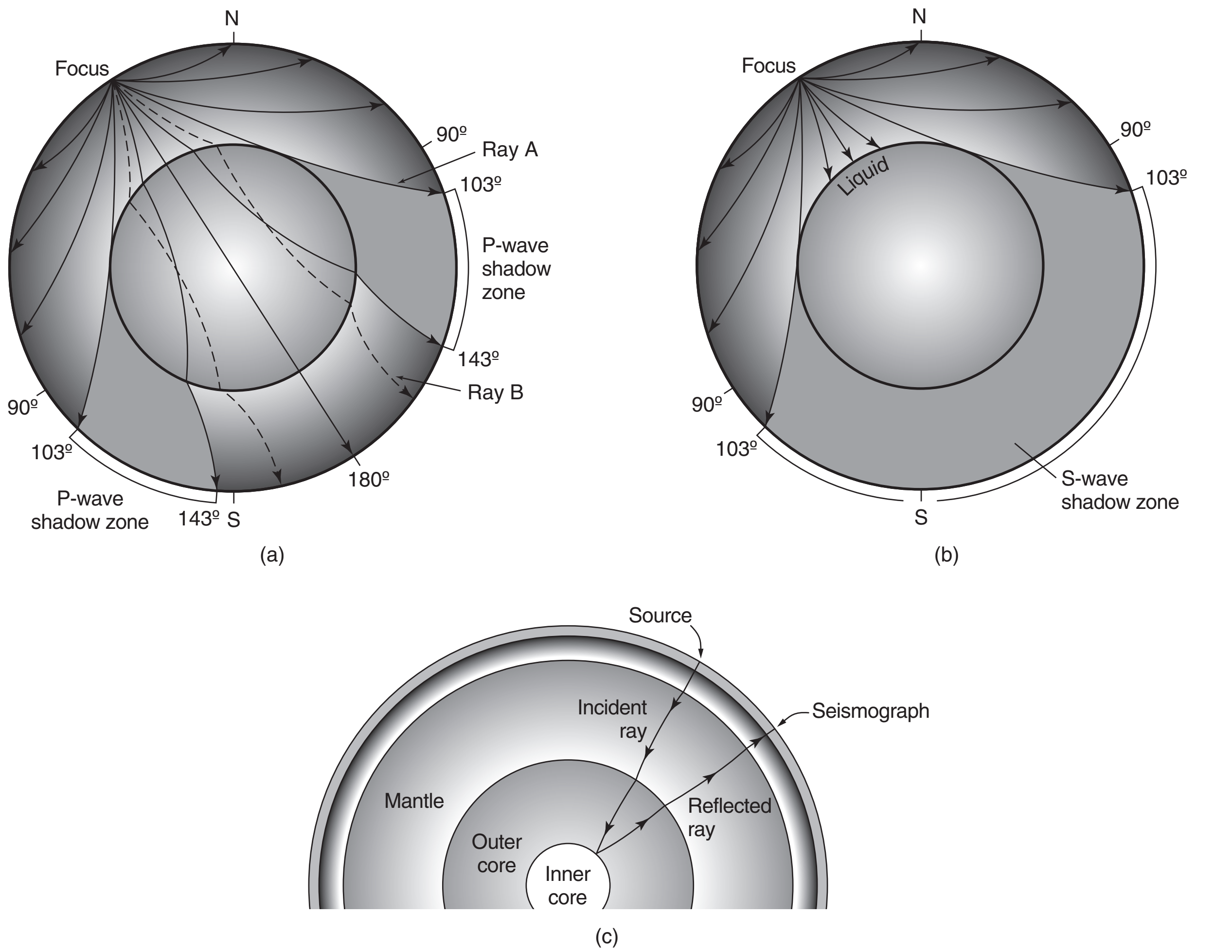
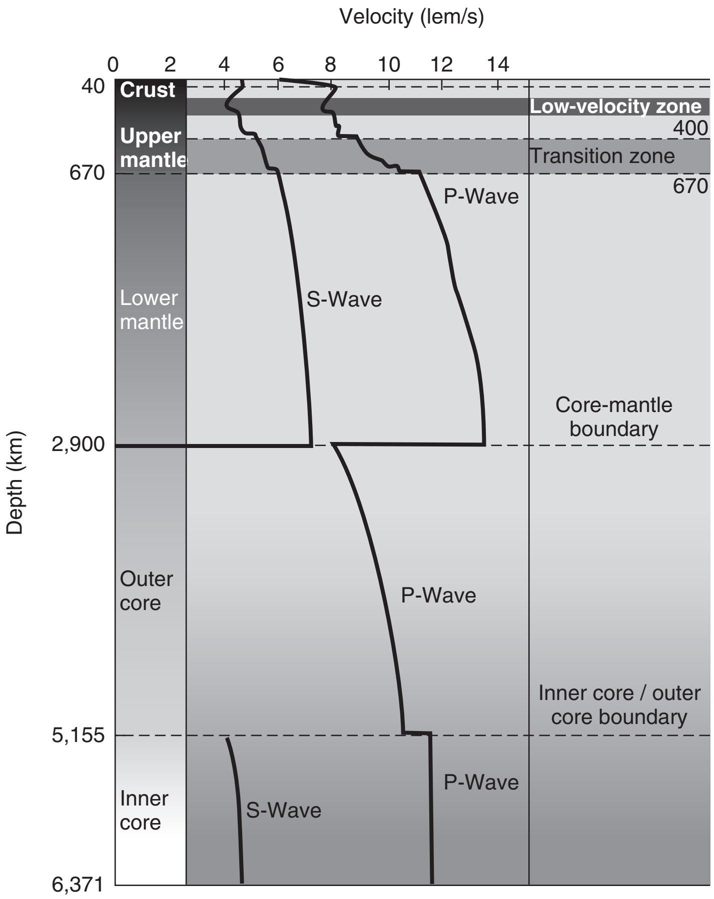


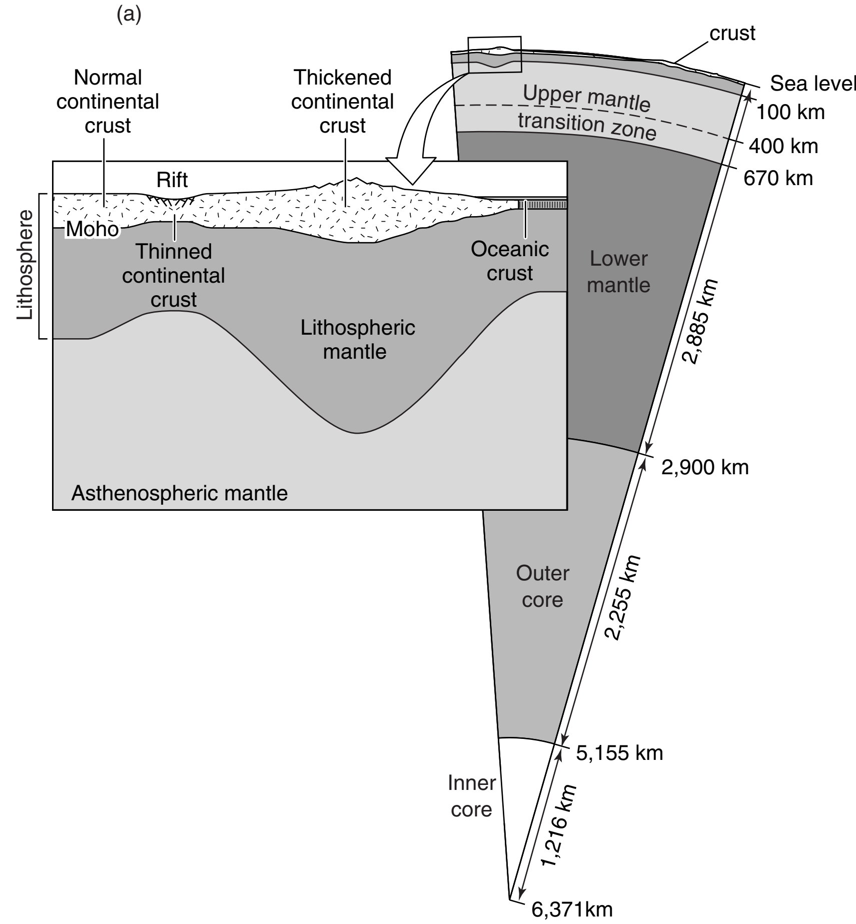
FIGURE 14.3 (a) Earthquake P-waves (compressional waves) generated at a focus travel into the Earth. We can represent the path that they follow by the use of "rays." A "ray" is an arrow drawn perpendicular to the wave front at a point. Note that waves bend or "refract" as they pass downwards. This happens because material properties of the Earth change with depth, so that wave velocity increases. If the change in material properties is gradual, the bending of a ray is gradual. At abrupt changes, the ray bends abruptly. P-waves bend abruptly at the mantle/core boundary. As a result, there is a P-wave shadow zone, where P-waves from a given earthquake do not reach the Earth's surface. The presence of this shadow zone proves the existence of the core. (b) S-waves (shear waves) also travel into the Earth. But shear waves cannot pass through a liquid, and thus cannot pass through the outer core. This creates a large S-wave shadow zone. It is the presence of this shadow zone that proves the existence of liquid in the core. (c) The existence of a solid inner core was deduced from studies that showed that earthquake waves bounce off a boundary within the core.

By studying the records of earthquakes worldwide, seismologists have been able to determine how long it takes seismic waves to pass through the Earth, depending on the path that the wave travels. From this data, they can calculate how seismic-wave velocity changes with depth. The data reveal that there are specific depths inside the Earth at which velocity abruptly changes and waves bend—these depths are called **seismic discontinuities** (Figure 14.3a and b). Seismic discontinuities divide the Earth's interior into distinct shells; within a shell, seismic wave velocity increases gradually, and waves bend gradually, but at the boundary between shells, wave velocity changes suddenly and the waves bend. The change in seismic velocity at a seismic discontinuity can be a consequence of a **compositional change,** meaning a change in the identity or proportion of atoms, and/or a **phase change,** meaning a rearrangement of atoms to form a new mineral structure. Phase changes occur because of changes in temperature and pressure.

Figure 14.4a illustrates an average **seismic velocity versus depth profile** for the Earth. Seismic discontinuities define layer boundaries, as shown in Figure 14.4b. Discontinuities led geologists to subdivide the earlier-mentioned major layers—crust, mantle, and core. We now recognize the following **seismically defined layers: crust, upper mantle, transition zone** (so named because it contains several small discontinuities), **lower mantle, outer core,** and **inner core.** Let's now look at the individual layers more closely.



(a)



(b)

FIGURE 14.4 (a) Graph of the variation in P-wave and S-wave velocities with depth in the Earth. The side of the graph shows the correlation of these boundaries with seismically defined layering in the Earth. (b) Diagram illustrating the internal layering of the Earth. Note the relationship between seismically defined layering and rheologic layering.

14.4 THE CRUST

The **crust** of the Earth refers to the outermost layer of our planet. All of the geologic structures that we have discussed in this book so far occur in rocks and sediments of the crust. Geologists distinguish between two fundamentally different types of crust on Earth: continental crust, which covers about 30% of the Earth's surface, and oceanic crust, which covers the remaining 70%. These types of crust differ from each other in terms of both thickness and composition.

The proportions of the two types of crust can be easily seen on a graph, called a **hypsometric curve,** that shows the percentage of Earth's surface area as a function of elevation or depth, relative to sea level (Figure 14.5a). Most continental crust lies between sea level and 1 km above sea level; mountains reach a maximum height of 8.5 km, and continental shelves (the submerged margins of continents) lie at depths of up to 600 m below sea level. Most oceanic crust lies at depths of between 3 km and 5 km below sea level. The greatest depth of the sea occurs in the Marianas Trench, whose floor lies over 11 km below sea level. The bottom of the crust is a seismic discontinuity called the **Moho.**[2] Interestingly, the elevation pattern of Earth is different from that of other bodies in the inner Solar System, as shown by histograms of elevation (Figure 14.5b). This distinction can be used as one piece of evidence that our planet's evolution has been different from that of the Moon, Mars, or Venus; Earth is probably the only planet whose surface has been affected by plate tectonics.

14.4.1 Oceanic Crust

Earth's oceanic crust is 6–10 km thick, and consists of mafic igneous rock overlain by a sedimentary blanket of varying thickness. Field studies of **ophiolites** (slices of oceanic crust emplaced on land by thrusting), laboratory studies of drill cores, and seismic-refraction studies (Table 14.1) indicate that oceanic crust has distinct layers. These layers, when first recognized in seismic-refraction profiles, were given the exciting names Layer 1, Layer 2a, Layer 2b, and Layer 3 (Figure 14.6). Layer 3, at the base, consists of **cumulate,** a rock formed from mafic (magnesium- and iron-rich) minerals that were the first to crystallize in a cooling magma and then settled to the bottom of the magma chamber. The cumulate is overlain in succession by a layer of **gabbro** (massive, coarse-grained mafic igneous rock), a layer of basaltic **sheeted dikes** (dikes that intrude dikes), a layer of **pillow basalt** (pillow-shaped blobs extruded into sea water), and a layer of **pelagic sediment** (the shells of plankton and particles of clay that settled like snow out of sea water). We'll explain how the distinct layers of crust form in Chapter 16.

Earth's ocean floor can be divided into four distinct **bathymetric provinces,** regions that lie within a given depth range and have a characteristic type of submarine landscape, defined in the following list. We indicate the plate-tectonic environments of these features, where appropriate (Figure 14.5c). These environments will be discussed later in the chapter.

- **Abyssal plains:** These are the broad, very flat, submarine plains of the ocean that lie at depths of between 3 km and 5 km. They are covered with a layer of pelagic (deep-sea) sediment.
- **Mid-ocean ridges:** These are long, submarine mountain ranges that rise about 2 km above the abyssal plains. Their crests, therefore, generally lie at depths of about 2–3 km. Mid-ocean ridges are roughly symmetric relative to a central axis, along which active submarine volcanism occurs. Mid-ocean ridges mark the presence of a divergent plate boundary, at which seafloor spreading occurs.
- **Oceanic trenches:** These are linear submarine troughs in which water depths range from 6 to 11 km. Trenches border an active volcanic arc and define the trace of a convergent plate boundary at which subduction occurs. The volcanic arc lies on the overriding plate.
- **Seamounts:** Seamounts are submarine mountains that are not part of mid-ocean ridges. They typically occur in chains continuous along their length with a chain of oceanic islands. The island at the end of the chain may be an active volcano. A seamount originates as a hot-spot volcanic island, formed above a mantle plume. When the volcano drifts off of the plume, it becomes extinct and sinks below sea level.
- **Guyots:** Guyots are flat-topped seamounts. The flat top may have been formed by the erosion of a seamount as the seamount became submerged, or it may be the relict of a coral reef that formed as the seamount became submerged.
- **Submarine plateaus:** These are broad regions where the ocean is anomalously shallow. Submarine plateaus probably form above large hot-spots.

14.4.2 Continental Crust

The continental crust differs in many ways from the oceanic crust (Table 14.2). To start with, the thickness

[2]Named for Andrija Mohorovičić, the seismologist who discovered it in 1909.

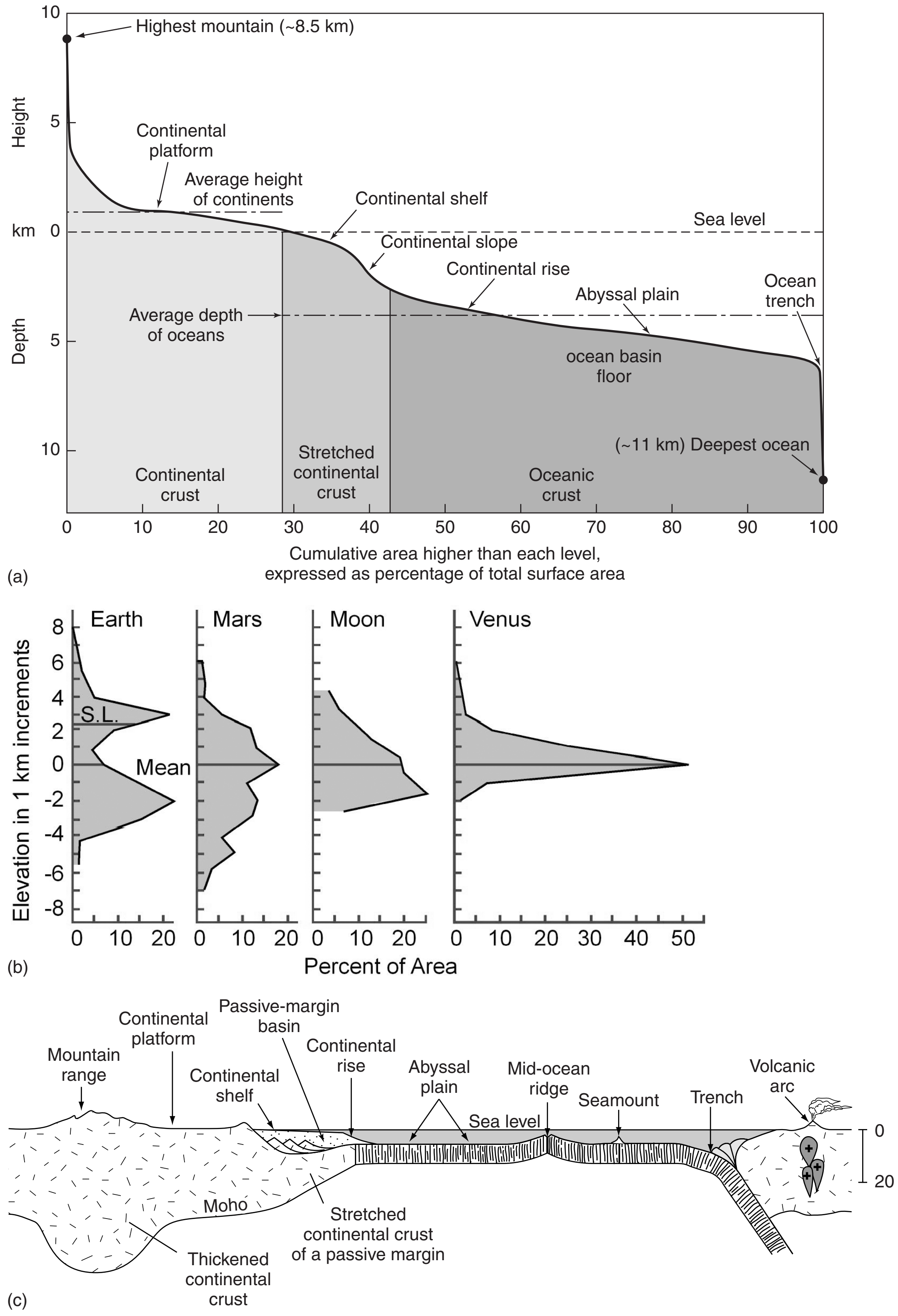


FIGURE 14.5 (a) Hypsometric curve of the Earth, showing elevation as a function of cumulative area. From this diagram, 29% of the Earth's surface lies above sea level; the deepest oceans and highest mountains comprise only a small fraction of the total area. The total surface area of the Earth is 510×10^6 km^2. Note that most of the continental shelf regions of the Earth coincide with passive margins, underlain by stretched continental crust. (b) Histograms that compare the elevation distribution of Earth, Mars, Venus, and Moon, showing that the bimodality of Earth's surface elevation (due to the presence of oceans and continents is not found on other planets). (c) Cross section of Earth's crust, showing various bathymetric features of the sea floor.

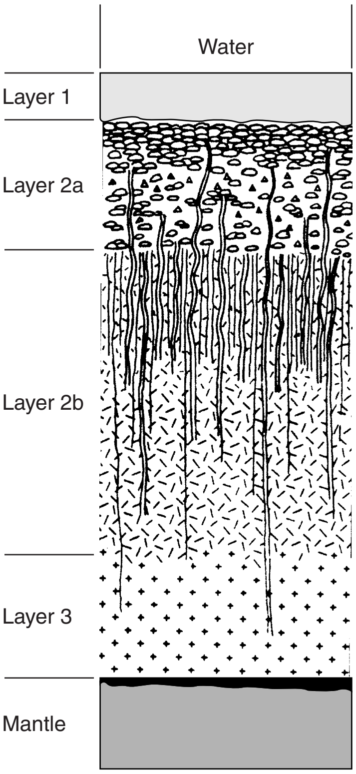


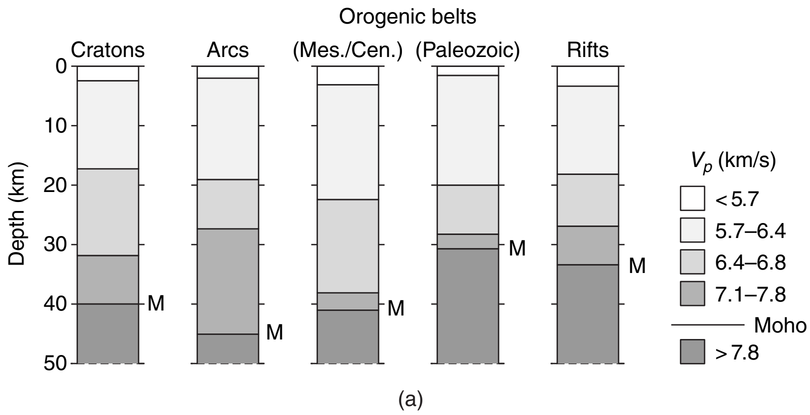
FIGURE 14.6 Columnar section of the oceanic crust and upper mantle. This section schematically represents the layers of the crust between the sea floor and the Moho.

of continental crust is, on average, four times that of oceanic crust. Thus, continental crust has an average thickness of about 35–40 km. In addition, continental crustal thickness is much more variable than oceanic crustal thickness. For example, beneath mountain belts formed where two continents collide and squash together, the crust can attain a thickness of up to 70 km, while beneath active rifts, where plates are being stretched and pulled apart, the crust thins to less than 25 km (Figure 14.7a). Next, if you were to calculate the average relative abundances of elements comprising continental crust, you would find that it has, overall, a silicic to intermediate composition (i.e., a composition comparable to that of granite to granodiorite). Thus, it is less mafic, overall, than oceanic crust. Finally, continental crust is very heterogeneous. In contrast, oceanic crust, which all forms in the same way at mid-ocean ridges, has the same overall structure everywhere on the planet. Specifically, continental crust consists of a great variety of different igneous, metamorphic, and sedimentary rocks formed at different times and in different tectonic settings during Earth history. The varieties of rock types have been deposited in succession, interleaved by faulting, corrugated by folding, or juxtaposed by intrusion (Figure 14.7b).

In the older part of continents can we find gross layering. Here, the upper continental crust has an average chemical composition that resembles that of granite

TABLE 14.2 DIFFERENCES BETWEEN OCEANIC AND CONTINENTAL CRUST

Composition	Continental crust has a mean composition that is less mafic than that of oceanic crust.
Mode of formation	Continental crust is an amalgamation of rock that originally formed at volcanic arcs or hot spots, and then subsequently passes through the rock cycle. Mountain building, erosion and sedimentation, and continued volcanism add to or change continental crust. Oceanic crust all forms at mid-ocean ridges by the process of seafloor spreading.
Thickness	Continental crust ranges between 25 km and 70 km in thickness. Most oceanic crust is between 6 km and 10 km thick. Thus, continental crust is thicker than oceanic crust.
Heterogeneity	Oceanic crust can all be subdivided into the same distinct layers, worldwide. Continental crust is very heterogeneous, reflecting its complex history and the fact that different regions of continental crust formed in different ways.
Age	Continental crust is buoyant relative to the upper mantle, and thus cannot be subducted. Thus, portions of the continental crust are very old (the oldest known crust is about 4 Ga). Most oceanic crust, gets carried back into the mantle during subduction, so there is no oceanic crust on Earth older than about 200 Ma, with the exception of the oceanic crust in ophiolites that have been emplaced and preserved on continents.
Moho	The Moho at the base of the oceanic crust is very sharp, suggesting that the boundary between crust and mantle is sharp. The continental Moho tends to be less distinct.



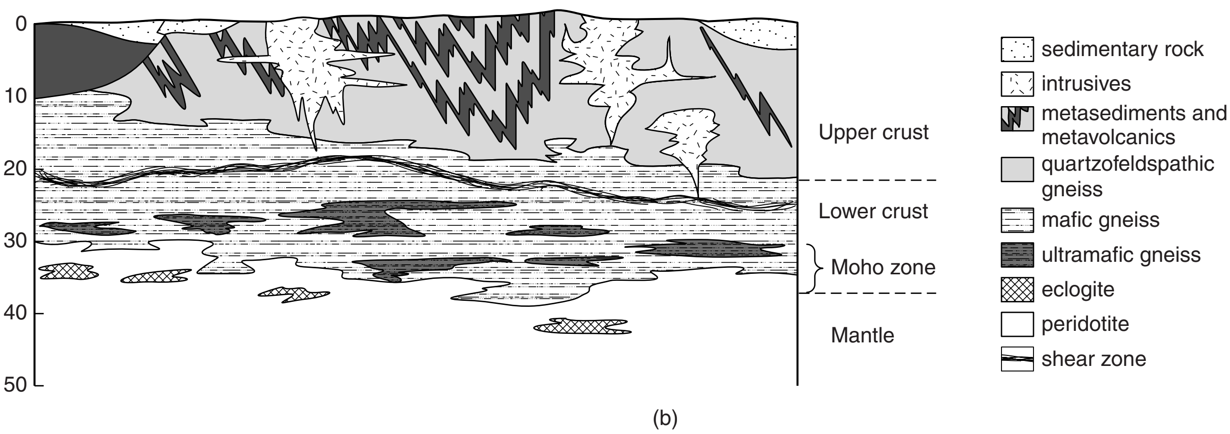


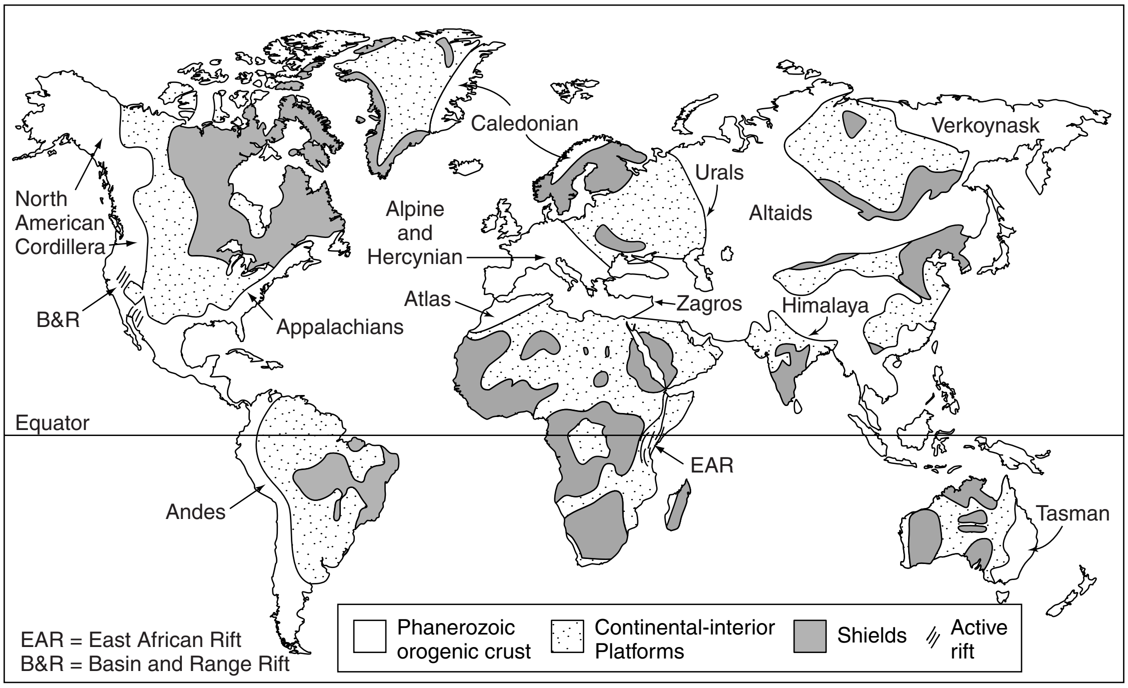
FIGURE 14.7 (a) Vertical sections through continental crust of different tectonic settings, based primarily on seismic data. The P-wave velocities are shown; M is Moho. Note the varying thickness of the highest velocity layer in the lower crust, which ranges from 20 km in arc regions to only a few kilometers in young orogenic belts. (b) Schematic cross section of the continental crust, illustrating its complexity.

(a felsic igneous rock) or granodiorite (an intermediate composition igneous rock), while the lower continental crust has an average chemical composition that resembles that of basalt (a mafic igneous rock). The boundary between upper crust and lower crust occurs at a depth of about 25 km. Contrasts in composition between upper and lower continental crust probably resulted, in part, from the differentiation of the continental crust during the Precambrian. Perhaps, partial melting of the lower crust created intermediate and silicic magmas which then rose to shallow levels before solidifying, leaving mafic rock behind. Mafic rock in the lower crust may also have formed from magma that formed by partial melting in the mantle, that rose and pooled at the base of the crust or intruded to form sills near the base of the crust. The process by which mafic rock gets added to the base of the crust is called **magmatic underplating.** Note that the compositional similarity between the lower continental crust and the oceanic crust does *not* mean that there is oceanic crust at the base of continents.

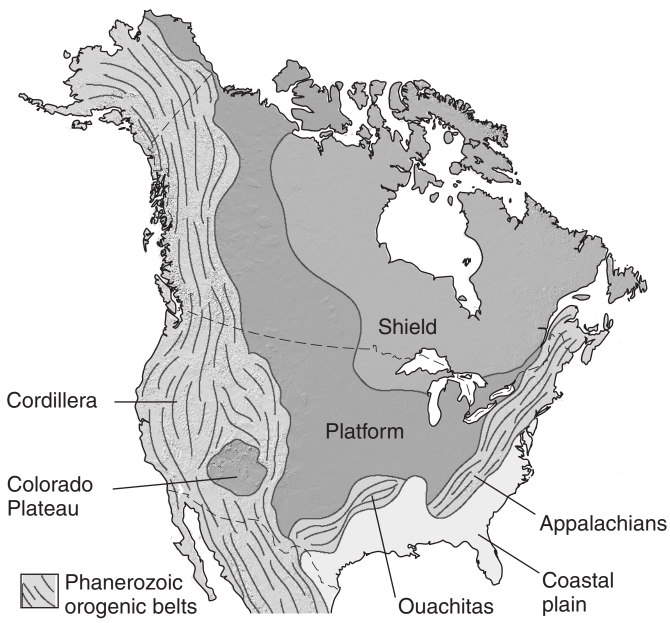
The heterogeneity of the continental crust reflects its long and complicated history. Most of the atoms forming the present continental crust were extracted from the Earth's mantle by partial melting and subsequent rise of magma in Archean and Paleoproterozoic time, between about 3.9 Ga and 2.0 Ga. Early continental crust grew by the amalgamation of volcanic arcs formed along convergent plate boundaries, as well as of oceanic plateaus formed above hot spots. Once formed, continental crust is sufficiently buoyant that it cannot be subducted, but it can be recycled through the stages of the rock cycle.

Because of its heterogeneity, geologists have found that it is useful to distinguish among several categories of continental crust (Figures 14.7a and 14.8a). Differences among these categories are based on: the age and type of rock making up the crust; the time when the crust was last involved in pervasive metamorphism and deformation; and the style of tectonism that has affected the crust.

Precambrian shields are broad regions of continents in which Precambrian rocks are presently exposed at the Earth's surface (Figure 14.8b). Most of these rocks are plutonic or metamorphic, but locally, shields contain exposures of Precambrian volcanic and sedimentary rocks. Some shields (e.g., the Canadian and Siberian shields) have relatively low elevations and subdued topography, but others (e.g., the Brazilian shield) attain elevations of up to 2 km above sea level and have been deeply disected by river erosion. Fault-bounded slices of Precambrian rocks incorporated in Phanerozoic mountain belts are not considered to be shields.



(a)

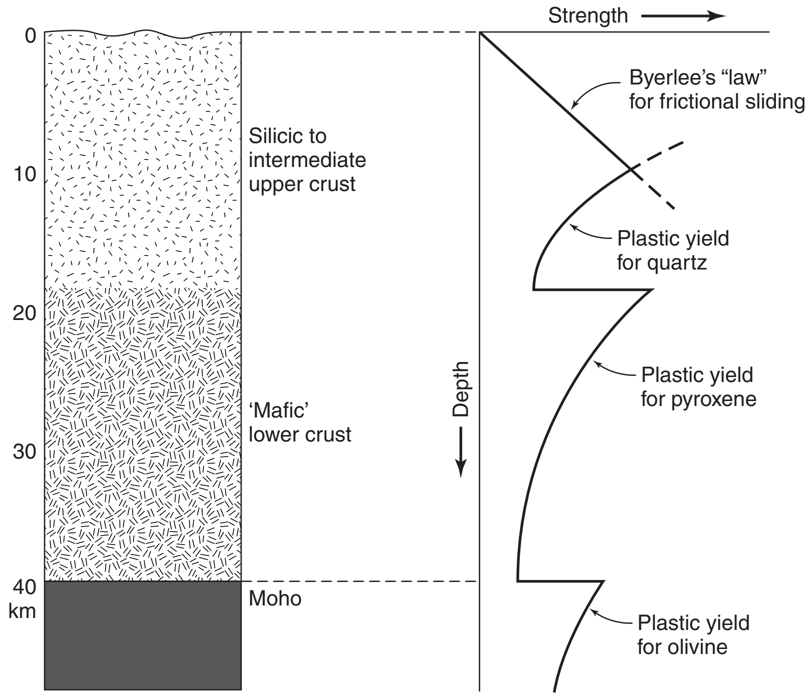


(b)

FIGURE 14.8 (a) Map of the continents, showing the distribution of different kinds of continental crust. Ancient rifts and details of mountain belts are not shown. (b) Detail of North America, illustrating the location of the platform and shield. The coastal plain is the region where the Appalachian Orogen has been buried by Mesozoic and Cenozoic strata.

Continental platforms are regions where Precambrian rocks are currently covered by a relatively thin veneer of unmetamorphosed and generally flat-lying latest Precambrian (Neoproterozoic) or Phanerozoic sedimentary strata. Geologists commonly refer to the rock below the sedimentary veneer as "basement" and the sedimentary veneer itself as "cover." Cover strata of platforms is, on average, less that 2 km thick, but in intracratonic basins, it may reach a maximum thickness of 5 to 7 km. The Great Plains region of the American Midwest is part of North America's continental platform (Figure 14.8b).

Cratons are the old, relatively stable portions of continental crust. In this context, we use the word "stable" to mean that cratonic crust has not been pervasively deformed or metamorphosed during at least the past one billion years. As such, cratons include both continental platforms and pre-Neoproterozoic Precam-



(c)

FIGURE 14.8 (continued) (c) Schematic graph illustrating the variation of strength with depth beneath the continental crust. Strength (measured in arbitrary units of stress) is controlled by frictional sliding in the upper crust, and by plastic deformation mechanisms below. The composition of the crust determines which mineral's flow law determines crustal strength.

brian shields (Figure 14.8b). Because of their age, cratons tend to be relatively cool (i.e., they are regions with low heat flow), and thus quite strong. Beneath cratons, the crust has a thickness of 35 to 40 km.

Younger orogenic belts are regions of continental crust that have been involved in orogeny during the Neoproterozoic or Phanerozoic. This means that they contain igneous rocks, metamorphic rocks, and geologic structures that are younger than about one billion years old. The crust in younger orogenic belts can be either "juvenile" (meaning that it formed by extraction of new melt from the mantle at the time of orogeny), or "recycled" (meaning that it is composed of older rock that was remelted, was intensely metamorphosed, or was transformed into sediment and then lithified at the time of orogeny). Geologists subdivide younger orogenic belts into two subcategories: **active or recently active orogens** (e.g., the Alps, the Andes, and the Himalays) are those in which topography and crustal thickness still reflect the consequences of mountain-building processes, while **inactive orogens** (e.g., the Appalachians, the Caledonides, and the Brasiliano/Pan African orogens in South America and Africa) are those in which tectonic activity ceased so long ago that erosion has exhumed deep levels of the orogen and the crust is not necessarily anomalously thick. Notably, in inactive orogens, current topography is not necessarily a relict of the uplift accompanying the original orogeny, but may be due to a much younger, independent uplift event. Because crust slowly cools through time, younger orogenic belts are warmer than cratons. Not surprsingly, heat flow in active orogens is greater than that of inactive orogens. Younger orogens generally have thicker crust than do cratons. In fact, beneath some active orogens (e.g., the Himalayan orogen), the crust reaches a thickness of 65 to 70 km.

Rifts are regions where the continental lithosphere has been stretched and thinned. As we will see, rifting leads to the formation of normal faults, formation of deep sedimentary basins, and volcanic activity. Geologists subdivide rifts into **active rifts,** where stretching continues today and active faulting is occurring (e.g., the Basin and Range of the western United States, and the East African Rift of Africa), and **inactive rifts,** where stretching ceased long ago (e.g., the Mesozoic rifts of the eastern United States). Crust in active rifts may be only 25 km thick, and is very warm—heat flow in active rifts may be many times that of cratons. Crustal thickness in inactive rifts is variable, as it reflects the history of tectonic events that the rift has endured subsequent to its original formation.

Passive margins are regions of continental crust that were stretched during a rifting event that succeeded in breaking apart a supercontinent to form two pieces that are now separated by an ocean basin. When rifting ceased and sea-floor spreading began, the relict of the rift became the tectonically inactive transition between the continent and the ocean basin (hence the name "passive margin"). As time passed, the stretched continental crust of passive margins slowly subsided (sank), and became buried by a very thick wedge of sediment derived from erosion of the adjacent continent or from the growth of reefs. The top surface of this sedimentary wedge forms a broad continental shelf (e.g., the region of relatively shallow ocean bordering the east coast of North America). Crustal thickness (including the wedge of sediment) and heat flow of a passive margin depend on the age of the margin. Older margins are thicker and cooler.

The strength of continental crust depends on its composition, thickness, and temperature. Temperature, as represented by heat flow, is particularly important because warmer crust is weaker than cooler crust. This relationship between temperature and strength simply reflects the fact that warmer rock is more ductile than cold rock. Continental crustal types can be ranked in order of relative strength. Active rifts are the weakest. The strength of younger orogenic belts varies greatly depending on the age of the belt, for their strength depends on their warmth. Specifically, Cenozoic orogens are warmer and weaker than Early Paleozoic orogens. In cratons, old shields are the strongest, while continental platforms are weaker, but both types of crust are stronger than younger orogenic belts. The strength of crust determines its behavior during tectonic activity. For example, the Cenozoic collision of India with Asia had little effect on the strong, old craton of India, but transformed southern Asia, an amalgamation of weak younger orogenic belts, into a very broad mobile belt. The interaction somewhat resembles a collision between an armored bank truck and a mountain of jelly. Similarly, it is much easier to rift apart a supercontinent along a recent orogen, than it is to rift apart an Archean craton.

Because the strength of continental crust varies with composition and temperature, strength varies with depth. The graph of Figure 14.8c illustrates this behavior. In the upper, cooler crust, strength is controlled by the frictional behavior of rocks as defined by Byerlee's law (see Chapter 6). Since frictional strength is linearly proportional to the normal stress across the sliding surface, strength increases with depth, following a straight-line relation. At greater depth, below the brittle–plastic transition, temperature and confining pressure become great enough that the rock becomes plastic. At these depths, the strength of rock is controlled by the plasticity of quartz, the dominant mineral of upper crustal rocks. Plastic strength decreases exponentially with increasing depth, because strength depends on temperature and temperature increases with depth. At the boundary between the upper and lower crust, rock composition changes and olivine becomes the dominant mineral. Olivine is stronger than quartz under the same conditions of pressure and temperature, so we predict an abrupt increase in strength at this boundary. Below the boundary, strength again decreases exponentially, due to the plastic behavior of olivine.

14.4.3 The Moho

The **Moho,** short for *Mohorovičić discontinuity,* is a seismic discontinuity marking the base of the crust. Above the Moho, seismic P-waves travel at about 6 km/s, while below the Moho, their velocities are more than 8 km/s (Figure 14.7a). Beneath the ocean floor, this discontinuity is very distinct and probably represents a very abrupt contact between two rock types (gabbro above and peridotite below). Beneath continents, however, the Moho is less distinct; it is a zone that is 2–6 km thick in which P-wave velocity discontinuously changes. Locally, continental Moho appears to coincide with a distinct reflector in seismic-reflection profiles, but this is not always the case. Possibly the Moho beneath continents represents different features in different locations. In some locations, it is the contact between mantle and crustal rock types, whereas elsewhere it may be a zone of sill-like mafic intrusions or underplated layers in the lower crust.

14.5 THE MANTLE

14.5.1 Internal Structure of the Mantle

The mantle comprises the portion of the Earth's interior between the crust and the core. Since the top of the mantle lies at a depth of about 7–70 km, depending on location, while its base lies at a depth of 2,900 km beneath the surface, the mantle contains most of the Earth's mass. In general, the mantle consists of very hot, but solid rock. However, at depths of about 100–200 km beneath the oceanic crust, the mantle has undergone a slight amount of partial melting. Here, about 2–4% of the rock has transformed into magma, and this magma occurs in very thin films along grain boundaries or in small pores between grains. Chemically, mantle rock has the composition of **peridotite,** meaning that it is an ultramafic igneous rock (i.e., it contains a very high proportion of magnesium and iron oxide, relative to silica).

Researchers have concluded that the chemical composition of the mantle is roughly uniform throughout. Nevertheless, the mantle can be subdivided into three distinct layers, delineated by seismic-velocity discontinuities. These discontinuities probably mark depths at which minerals making up mantle peridotite undergo abrupt phase changes. At shallow depths, mantle peridotite contains olivine, but at greater depths, where pressures are greater, the olivine lattice becomes unstable, and atoms rearrange to form a different, more compact lattice. The resulting mineral is called β-phase. The phase change from olivine to β-phase takes place without affecting the overall chemical composition. The three layers of the mantle, from top to bottom, are as follows:

- **Upper mantle.** This is the shell between the Moho and a depth of about 400 km. Beneath ocean basins, the interval between about 100 km and 200 km has anomalously low seismic velocities; this interval is known as the **low-velocity zone.** As noted earlier, the slowness of seismic waves in the low-velocity zone may be due to the presence of partial melt; 2–4% of the rock occurs as magma in thin films along grain surfaces, or in small pores between grains. Note that the low-velocity zone constitutes only part of the upper mantle and that it probably does not exist beneath continents.
- **Transition zone.** This is the interval between a depth of 400 km and 670 km. Within this interval, we observe several abrupt jumps in seismic velocity, probably due to a succession of phase changes in mantle minerals.
- **Lower mantle.** This is the interval between a depth of 670 km and a depth of 2,900 km. Here, temperature, pressure, and seismic velocity gradually increase with depth.

As we noted above, the mantle consists entirely of solid rock, except in the low-velocity zone. But mantle rock is so hot that it behaves plastically and can flow very slowly (at rates of no more than a few centimeters per year). Given the slow rate of movement, it would take a mass of rock about 50–100 million years to rise from the base to the top of the mantle

Because of its ability to flow, the mantle slowly undergoes **convection** (Table 14.3). Convection is a mode of heat transfer during which hot material rises, while cold material sinks. (You can see convection occurring as you warm soup on a stove.) Convective movement takes place in the mantle because heat from the core warms the base of the mantle. Warm rock is less dense than cool rock, and thus feels a buoyant force pushing it upwards. In other words, warmer rock has "positive buoyancy" when imbedded in cooler rock. In the mantle, this buoyant force exceeds the strength of plastic peridotite, and thus buoyant rock originating deep in the mantle can rise. As it does so, it pushes aside other rock in its path, like a block of wood rising through water pushes aside the water in its path. Meanwhile, cold rock, at the top of the mantle, is denser than its surroundings, has "negative buoyancy" and sinks, like an anchor sinking through water. Geologists do not know for sure whether the upper mantle and lower mantle convect independently, and remain as separate chemical reservoirs, or if they mix entirely or at least partially during convection.

Because of convection, the onion-like image of the mantle that we provided above is actually an oversimplification. **Seismic tomography** (a technique for generating a three dimensional image of the Earth's interior; Table 14.1 and Figure 14.9) suggests that, in fact, the mantle is heterogeneous within

TABLE 14.3	TYPES OF HEAT TRANSPORT INSIDE THE EARTH
Conduction	This phenomenon occurs when you place one end of an iron bar in a fire; heat gradually moves along the bar so that the other end eventually gets hot too. We say that heat flows along the bar by conduction. Note that iron atoms do not physically move from the hot end of the bar to the cool end. Rather, what happens is that the atoms nearest the fire, when heated, vibrate faster, and this vibration, in turn, causes adjacent atoms further from the fire to vibrate faster, and so on, until the whole bar becomes warm.
Convection	This phenomenon occurs when you place a pot of soup on a hot stove. Conduction through the base of the pot causes the soup at the bottom to heat up. The heated soup becomes less dense than the overlying (cold) soup; this density inversion is unstable in a gravity field. Consequently, the hot soup starts to rise, to be replaced by cold soup that sinks. Thus, convection is driven by density gradients that generate buoyancy forces and occurs by physical flow of hot material. Convection occurs when (1) the rate at which heat is added at the bottom exceeds the rate at which heat can be conducted upward through the layer, and (2) the material that gets heated is able to flow.
Advection	This phenomenon occurs when water from a boiler flows through pipes and the pipes heat up. The water carries heat with it, and the heat conducts from the water into the metal. Thus, advection is the process by which a moving fluid brings heat into a solid or removes heat from a solid. In the Earth, heating by advection occurs where hot water or hot magma passes through fractures in rock, and heats the surrounding rock. Cooling by advection occurs where cold seawater sinks into the oceanic crust, absorbs heat, and then rises, carrying the heat back to the sea.

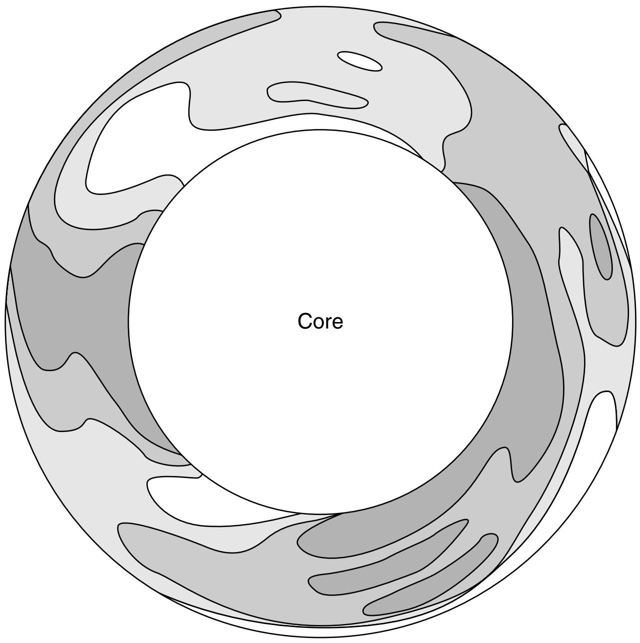


FIGURE 14.9 Seismic tomographic image of the mantle. The shading indicates differences in the velocity of seismic waves, with the region of highest velocity (coldest material) indicated by the darkest color and the region of lowest velocity (hotter material) by white. Note that, in places, these patterns suggest mantle convection, in which relatively hot material rises from the core-mantle boundary region to the upper mantle.

the layers. Specifically, tomographic studies reveal that, at a given depth, the mantle contains regions where seismic velocities are a little faster, and regions where seismic velocities are a little slower, than an average value. These variations primarily reflect variations in temperature (waves slow down in hotter, softer mantle). An overall tomographic image shows blobs and swirls of faster-transmitting mantle interspersed with zones of slower-transmitting mantle. The slower-transmitting mantle is warmer, and less dense, and thus is rising, while the faster-transmitting mantle is colder, and denser, and thus is sinking.

14.5.2 Mantle Plumes

A map of active volcanoes on the Earth reveals that most volcanoes occur in chains (which, as we see later in the chapter, occur along the boundaries between plates), but that some volcanoes occur in isolation. For example, several active, or recently active, volcanoes comprise the Cascades chain in Oregon and Washington, but the big island of Hawaii erupts by itself in the middle of the Pacific Ocean, far distant from any other volcano. Geologists refer to isolated volcanic sites as hot spots. There are about 100 **hot spots** currently active on the surface of the Earth.

Geologists speculate that hot-spot volcanoes form above **mantle plumes,** columns of hot mantle rising from just above the core-mantle boundary. According to this model, mantle plumes form because heat conducting out of the core warms the base of the mantle, creating a particularly hot, positively buoyant layer.[3] At various locations, the surface of this layer bulges up, and a dome of hot rock begins to rise diapirically. This dome evolves into a column or "plume." Beneath a hot spot volcano, a plume has risen all the way to the base of the lithosphere. Because of the decompression that accompanies this rise, the peridotite at the top of the plume partially melts, producing basaltic magma. This magma eventually rises through the lithosphere to erupt at the Earth's surface.

14.6 THE CORE

The core of the Earth has a diameter of about 3,481 km; thus, the core is about the size of the Moon. Because of its great density, geologists have concluded that the core consists of iron alloy. As indicated in Figure 14.3a, a seismic discontinuity divides the core into two parts. Note that seismic P-waves (compressional waves) can travel through both the outer core and the inner core, while seismic S-waves (shear waves) travel through the inner core but not through the outer core. Shear waves cannot travel through a liquid, so this observation means that the outer core consists of liquid (molten) iron alloy, while the inner core consists of solid iron alloy. Convective flow of iron alloy in the outer core probably generates Earth's magnetic field. Of note, recent studies indicate that the inner core does not rotate with the same velocity as the mantle. Rather, the inner core spins slightly faster. Similar studies show that the inner core is anisotropic, so that seismic waves travel at different speeds in different directions through the inner core.

14.7 DEFINING EARTH LAYERS BASED ON RHEOLOGIC BEHAVIOR

In the early years of the twentieth century, long before the discovery of plate tectonics, geologists noticed that the thickness of the crust was not uniform, and that

[3]Seismologists refer to this thin hot layer as the D″ layer (D″ is pronounced "D double prime").

where crust was thicker, its top surface lay at higher elevations. This behavior reminded geologists of a bathtub experiment in which wood blocks of different thickness were placed in water—the surface of the thicker blocks lay at a higher elevation than the surface of thinner blocks. Could the crust (or the crust together with the topmost layer of mantle) be "floating" on a weaker layer of mantle below? If so, then the mantle, below a certain depth, would have to be able to respond to stress by flowing. Geologists also noticed that when a large glacier grew on the surface of the Earth, it pushed the surface of the Earth down, creating a broad dimple that was wider than the load itself. (To picture this dimple, imagine the shape of the indentation that you make when you stand on the surface of a trampoline or a mattress.) Could the outermost layer of the Earth respond to loads by flexing, like a stiff sheet of rubber?

To explain the above phenomena, geologists realized that we needed to look at earth layering, at least for the outermost several hundred kilometers, by considering the way layers respond to stress. In other words, we had to consider **rheologic layering** (Figure 14.10a). By doing this, we can divide the outer several hundred kilometers into two layers based on the way that layers respond to stress. This perspective is different from that provided by studying seismic discontinuities in the Earth.

We refer to the outer, relatively rigid shell of the Earth that responds to stress by bending or flexing, as the **lithosphere.** The lithosphere, which consists of the crust and outermost mantle, overlies an interval of the mantle that responds to stress by plastically flowing. This plastic layer is called the **asthenosphere.** When plate tectonics theory came along in the 1960s, it proved convenient to picture the plates as large pieces of lithosphere. In fact the term "lithospheric plate" has become standard geologic jargon.[4] The plates could move over the asthenosphere because the asthenosphere is weak and plastic.

14.7.1 The Lithosphere

The **lithosphere,** derived from the Greek word "lithos" for rock (implying that it has strength, or resistance to stress), is the uppermost rheologic layer of the Earth. As noted above, lithosphere consists of the crust and the uppermost (i.e., coolest and strongest) part of the mantle; the mantle part of the lithosphere is called the **lithospheric mantle.** As we've noted already, the lithosphere can be distinguished rheologically from the underlying layer, the asthenosphere, because, overall, the lithosphere behaves rigidly on geologic timescales. If you place a load on it, the lithosphere either supports the load or bends—it does not simply flow out of the way. In technical terms, we say that lithosphere has "flexural rigidity."

To picture what we mean by flexural rigidity, imagine a stiff rubber sheet resting on a layer of honey (Figure 14.11a). The sheet has flexural rigidity, so if you place an empty can (a small load) on the sheet, the sheet supports it, and if you place a concrete block (a large load) on the sheet, the sheet bends. If we were to do the same experiment with a material that does not have flexural rigidity, the results would be different. For example, if we place a can directly on the surface of a pool of honey, a material without flexural rigidity, the can sinks in a little and then floats, while if we place the concrete block in the honey, the block sinks (Figure 14.11b). The honey is able to flow out of the way of the block as the block sinks and can then flow over the top of the block as the block passes. In sum, **flexural rigidity** is the resistance to bending (flexure) of a material. A steel beam has a relatively high flexural rigidity, but a sheet of rubber has low flexural rigidity. Totally nonelastic materials, like honey or lava, have no flexural rigidity at all. Not all lithosphere on the Earth has the same flexural rigidity. Old, cold cratons have greater flexural rigidity than younger and warmer orogens.

Significantly, the rheologic behavior of the lithosphere affects the way in which heat can be transported through it. Because the lithosphere cannot flow easily, heat moves through the lithosphere only by conduction or advection (see Table 14.3), not by convection. In contrast, as we see later, heat is transported in the asthenosphere primarily by convection.

As we noted above, the lithosphere includes all of the crust and the uppermost part of the mantle. For simplicity, the base of the lithosphere can be defined by an isotherm, meaning an imaginary surface on which all points have the same temperature. Geologists often define the 1280° C isotherm in the mantle as the base of the lithosphere, because at approximately this temperature, olivine, the dominant silicate mineral in mantle peridotite, becomes very weak; this weakness happens because dislocation glide and climb become efficient deformation mechanisms at high temperatures.

[4]Oceanic plates indeed seem to be composed only of lithosphere. They are decoupled from the asthenosphere at the very weak low-velocity zone. However, recent work suggests that continental plates have roots that extend down into the asthenosphere, in that some of the asthenosphere, perhaps down to a depth of 250 km, moves with the continental lithosphere when a plate moves. This thicker entity—lithosphere plus coupled asthenosphere—has been called the "tectosphere." The coupled asthenosphere may have subtle chemical differences that make it buoyant relative to surrounding asthenosphere. For simplicity in this book, we will consider plates to be composed of only lithosphere.

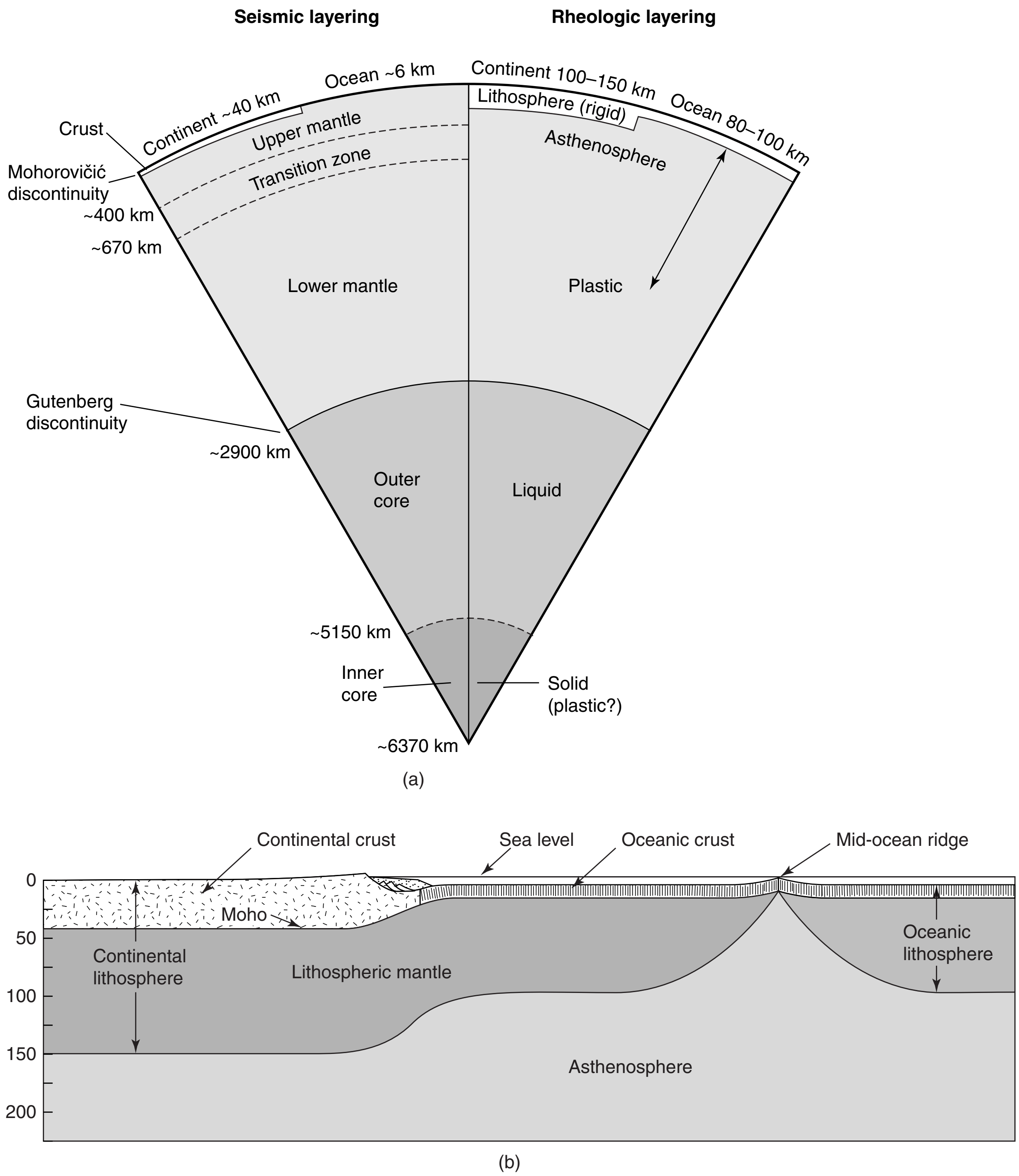


FIGURE 14.10 (a) Comparison of seismic layering and rheologic layering of the Earth. (b) Cross section of the lithosphere showing variations in lithosphere thickness.

In other words, we can view the boundary between the lithosphere and underlying asthenosphere as a **thermal boundary.** In contrast, the boundary between the crust and the upper mantle (which collectively make up the lithosphere) is a **compositional boundary,** meaning it is due to a change in chemical makeup.

The depth of the 1280° C isotherm is not fixed in the Earth, but varies depending on the thermal structure of the underlying mantle and on the duration of time that the overlying lithosphere has had to cool. Thus, the base of the lithosphere is not at a fixed depth everywhere around the Earth. Moreover, at a given location, the depth of the base of the lithosphere can change with time if the region is heated or cooled. For example, directly beneath the axis of mid-ocean ridges, the lithosphere is very thin, because the 1280° C isotherm rises almost to the base of the newly formed crust. Yet beneath the oceanic abyssal plains, oceanic lithosphere

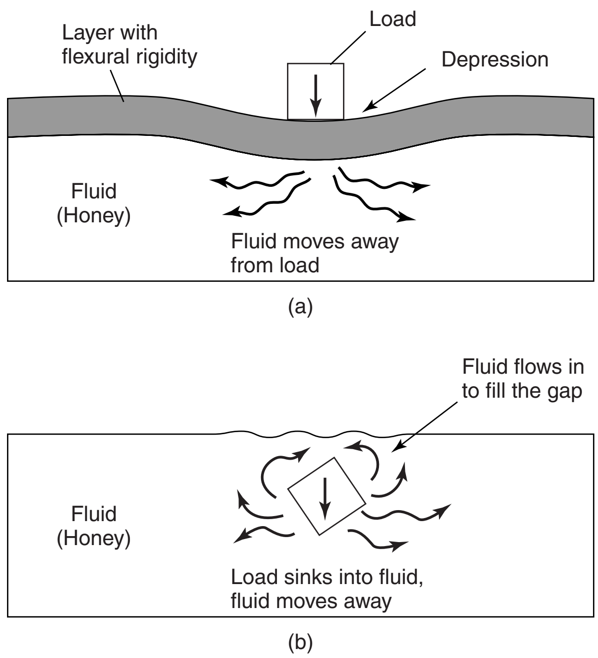


FIGURE 14.11 The concept of flexural rigidity. (a) A sheet of rubber has flexural rigidity and bends when a load is placed on it. The underlying honey flows out of the way. (b) A load placed in a material with no flexural rigidity simply sinks into it, if the load has negative buoyancy.

is old and has had time to cool; here it reaches a thickness of about 100 km. In other words, the lithospheric mantle of an oceanic plate thickens with time as seafloor spreading occurs, and the plate moves away from a mid-ocean ridge; this thickening occurs as heat conducts upward through the lithosphere to the surface of the Earth. Beneath continental cratons, the lithosphere is very ancient and may be more than 150 km thick.

14.7.2 The Asthenosphere

The **asthenosphere,**[5] which is the layer of the mantle that underlies the lithosphere, is so warm and plastic that it behaves somewhat like a viscous fluid. It does not have flexural rigidity over geologic time scales. (Though over very short timescales, it behaves elastically, for it can transmit seismic waves.) Thus, if a load is placed on the asthenosphere, the load sinks as the asthenosphere in the path of the load flows out of the way (Figure 14.11b). Because the asthenosphere has no flexural rigidity, lateral density differences that result from variations in temperature and/or composition can cause the asthenosphere to flow. Specifically, warmer and less dense parts of the asthenosphere rise, while denser parts sink; in other words, heat transfer in the asthenosphere occurs by convection, as we've already noted (Table 14.3). Keep in mind that even though the asthenosphere is able to flow, it is almost entirely composed of solid rock—it should *not* be viewed as a subterranean sea of magma. Indeed, the only magma in the asthenosphere occurs either in the low-velocity zone, where a slight amount of melt is found as films on the surface of grains, or in regions beneath volcanic provinces, where blobs of magma accumulate and rise. Plastic flow of the asthenosphere takes place largely by dislocation movement and diffusion.

As we've also noted, the top of the asthenosphere can be defined approximately by the 1280° C isotherm, for above this thermal boundary, peridotite is cool enough to behave rigidly, while below this boundary, peridotite is warm enough to behave plastically. There appears to be no significant contrast in chemical composition across this boundary. Defining the base of the asthenosphere is a matter of semantics, because all mantle below the lithosphere is soft enough to flow plastically. For convenience, some geoscientists consider the asthenosphere to be equivalent to the low-velocity zone in the upper portion of the upper mantle. Others, equate it with the interval of mantle between the base of the lithosphere and the top of the transition zone. Still others equate it with the interval between the base of the lithosphere and the top of the lower mantle, or even with all the mantle below the lithosphere. There really isn't a consensus on this issue.

14.7.3 Isostasy

Now that we have addressed the concept of lithosphere and asthenosphere, we can consider the principle of isostasy (or, simply, "isostasy"), which is an application of Archimedes' law of buoyancy to the Earth. Knowledge of isostasy will help you to understand the elevation of mountain ranges and the nature of gravity anomalies. To discuss isostasy, we must first review Archimedes' law.

Archimedes' law states that when you place a block of wood in a bathtub full of water, the block sinks until the mass of the water displaced by the block is equal to the mass of the whole block (Figure 14.12a). Since wood is less dense than the water, some of the block protrudes above the water, just like an iceberg protrudes above the sea. When you place two wood blocks of different thicknesses into the water, the surface of the thicker block floats higher than the surface of the thinner block, yet the proportion of the thick block above the water is the same as the proportion of the

[5]The term is derived from the Greek *asthenes,* meaning "weak."

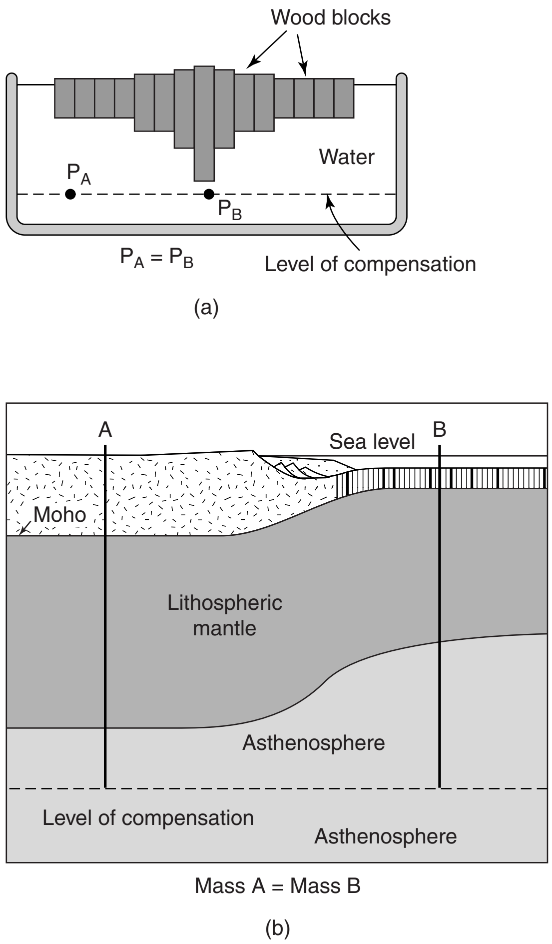


FIGURE 14.12 The concept of isostasy. (a) Blocks of wood of different thicknesses float at different elevations when placed in water. Therefore, the pressure at point A is the same as the pressure at point B. (b) In the Earth, isostasy requires that the mass of a column drilled down to the level of compensation at A equals the mass of a column drilled at B, if isostatic equilibrium exists at both locations.

thin block above the water. Thus, the base of the thick block lies at a greater depth than the base of the thin block. Now, imagine that you put two wood blocks of the same thickness but of different density in the water; this will be the case if one block is made of oak and the other is made of pine. The dense oak block floats lower than the less-dense pine block does. Note that, in the experiment, the pressure in the water at the base of the tub is the same regardless of which block floats above. Also, if you push down on or pull up on the surface of a block, it will no longer float at its proper depth.

If we make an analogy between the real Earth and our bathtub experiment, the lithosphere plays the role of the wooden blocks and the asthenosphere plays the role of the water. For a given thickness of lithosphere, the surface of more buoyant lithosphere floats higher than the surface of less buoyant lithosphere, if the lithosphere is free to float. Further, the pressure in the asthenosphere, at a depth well below the base of the lithosphere, is the same regardless of thickness and/or density of the lithosphere floating above (if the lithosphere is floating at the proper depth). We call a depth in the asthenosphere at which the pressure is the same, regardless of location, a **depth of compensation.**

With this image of floating lithosphere in mind, we can now state the **principle of isostasy** more formally as follows: When free to move vertically, lithosphere floats at an appropriate level in the asthenosphere so that the pressure at a depth of compensation in the asthenosphere well below the base of the lithosphere is the same. Where this condition is met, we say that the lithosphere is "isostatically compensated" or in "isostatic equilibrium."

Another way to picture isostatic equilibrium is as follows. If a location in the ocean lithosphere and a location in the continental lithosphere are both isostatically compensated, then a column from the Earth's surface to the depth of compensation at the ocean location has the same mass as a column of the same diameter to the same depth in the continental location (Figure 14.12b). Ocean basins exist because ocean crust is denser and thinner than continental crust, and thus ocean lithosphere sinks deeper into the asthenosphere than does continental lithosphere. Low-density water fills the space between the surface of the oceanic crust and the surface of the Earth.

Note that with isostasy in mind, we see that changing the relative proportions of crust and mantle within the lithosphere will change the depth to which the lithosphere sinks and thus will change the elevation of the lithosphere's surface. This happens because crustal rocks are less dense than mantle rocks. For example, if we increase the proportion of buoyant crust (by thickening the crust beneath a mountain range or by underplating magma to the base of crust), the surface of the lithosphere lies higher, and if we remove dense lithospheric mantle from the base of the plate, the plate rises.

If the lithosphere doesn't float at an appropriate depth, we say that the lithosphere is "uncompensated." Uncompensated lithosphere may occur, for example, where a relatively buoyant piece of lithosphere lies embedded within a broad region of less buoyant lithosphere. Because of its flexural rigidity, the surrounding lithosphere can hold the buoyant piece down at a level below the level that it would float to if unimpeded. The presence of uncompensated lithosphere causes gravity anomalies. Positive anomalies (gravitational pull is greater than expected) occur where there is excess mass, while negative anomalies (gravitational pull is less than expected) occur where there is too little mass.

14.8 THE TENETS OF PLATE TECTONICS THEORY

While lying in a hospital room, recovering from wounds he received in World War I, Alfred Wegener (1880–1930), a German meteorologist and geologist, pondered the history of the Earth. He wondered why the eastern coastlines of North and South America looked like they could cuddle snugly against the western coastlines of Europe and Africa. He wondered why glacial tills of Late Paleozoic age crop out in India and Australia, land masses that now lie close to the equator. And he wondered how a species of land-dwelling lizard could have evolved at the same time on different continents now separated from one another by a vast ocean. Eventually, Wegener realized that all these phenomena made sense if, in the past, the continents fit together like pieces of a jigsaw puzzle into one "supercontinent," a vast landmass that he dubbed **Pangaea.** According to Wegener, the supercontinent of Pangaea broke apart in the Mesozoic to form separate, smaller continents that have since moved to new locations on the Earth's surface. Wegener referred to this movement as **continental drift.** Before the breakup of Pangaea, Wegener speculated, the Atlantic Ocean didn't exist, so lizards could easily have walked from South America to Africa to Australia without getting their feet wet, and land that now lies in subtropical latitudes instead lay near the South Pole where it could have been glaciated.

Though Wegener's proposal that continents drift across Earth's surface seemed to explain many geologic phenomena, the idea did not gain favor with most geoscientists of the day because Wegener (who died while sledging across Greenland in 1931) could not provide an adequate explanation of *why* continents moved. Such an explanation did not appear until 1960, when an American geologist named Harry Hess circulated a manuscript in which he proposed a process known as **seafloor spreading** (a concept also credited to Robert S. Dietz). During seafloor spreading, new ocean floor forms along the axis of a submarine mountain range called a mid-ocean ridge. Once formed, the new ocean floor moves away from the ridge axis. Two continents drift apart when seafloor spreading causes the ocean basin between them to grow wider.

In the 1960s, researchers from around the world rushed to explore the implications of the seafloor spreading hypothesis and to reexamine the phenomenon of continental drift. The result of this work led to a broad set of ideas, which together comprise the **theory of plate tectonics**[6] (or simply "plate tectonics"). According to this theory, the lithosphere, Earth's relatively rigid outer shell, consists of discrete pieces, called **lithosphere plates,** or simply **plates,** which move relative to one another. Continental drift and seafloor spreading are manifestations of plate motion.

Plate tectonics is a **geotectonic theory.** It is a comprehensive set of ideas that explains the development of regional geologic features, such as the distinction between oceans and continents, the origin of mountain belts, and the distribution of earthquakes, volcanoes, and rock types. Acceptance of plate tectonics represented a revolution in geology, for it led to the inevitable conclusion that Earth's surface is mobile—the map of the planet constantly changes (though very slowly).[7]

A plate can be viewed geometrically as a cap on the surface of a sphere. The border between two adjacent plates is a **plate boundary.** During plate movement, the **plate interior** (the region away from the plate boundary) stays relatively coherent and undeformed. Thus, most plate movement is accommodated by deformation along plate boundaries, and this deformation generates earthquakes. A map of earthquake epicenters defines seismic belts, and these belts define the locations of plate boundaries (Figure 14.13).

A closer look at the map of earthquake-epicenter locations shows that not all plate boundaries are sharp lines. Along some, such as the boundary that occurs between India and Asia, and the boundary that occurs between parts of western North America and the Pacific Plate, earthquakes scatter over a fairly broad area. Such **diffuse plate boundaries** typically occur where plate boundaries lie within continental crust, for the quartz-rich continental crust is relatively weak and

[6]Recall that a "hypothesis" is simply a reasonable idea that has the potential to explain observations. A "theory" is an idea that has been rigorously tested and has not yet failed to explain relevant observations. Nevertheless, a theory could someday be proven wrong.

[7]Prior to the proposal of plate tectonics theory, most geologists had a "fixist" view of the Earth, in which continents were fixed in position through geologic time. In this context, geologists viewed mountain building to be a consequence predominantly of vertical motions. Pre–plate tectonics ideas to explain mountain building included: (1) The **geosyncline hypothesis,** which stated that mountain belts formed out of the deep, elongate sedimentary basins (known as "geosynclines") that formed along the margins of continents. According to this hypothesis, mountain building happened when the floor of a geosyncline sank deep enough for sediment to melt; the resulting magma rose and in the process deformed and metamorphosed surrounding rock; (2) The **contracting Earth hypothesis,** which stated that mountains formed when the Earth cooled, shrank, and wrinkled, much like a baked apple removed from the oven. Both ideas have been thoroughly discredited.

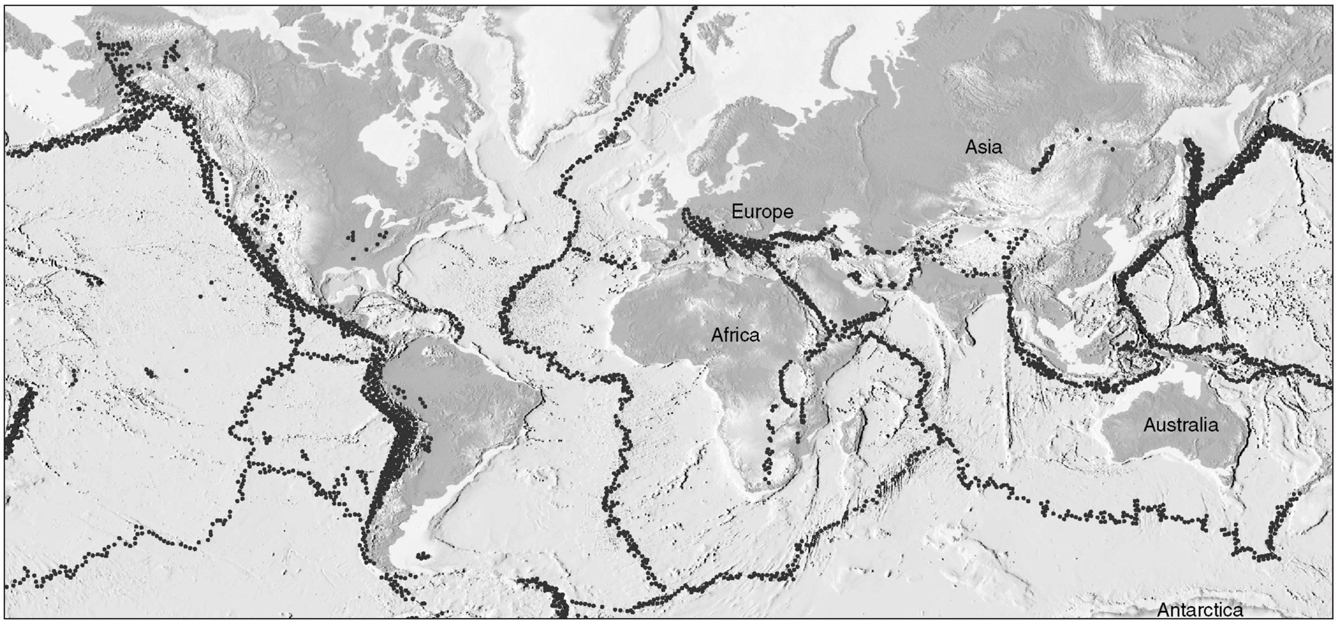
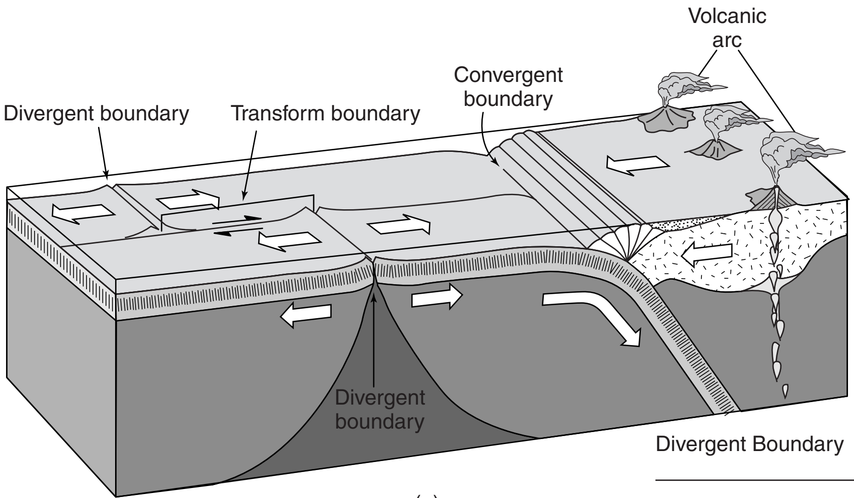


FIGURE 14.13 A map of earthquake epicenters (dots) outlines the locations of plate boundaries. Intraplate earthquakes occur away from plates.

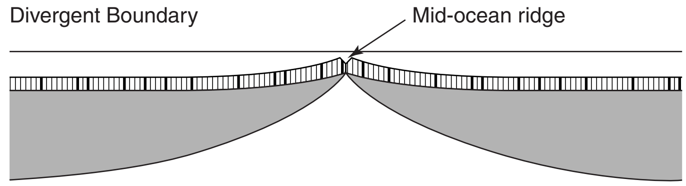
continental crust contains many preexisting faults, so the deformation is not confined to a narrow zone. Some seismic activity and deformation does occur entirely away from a plate boundary. Such activity, called **plate-interior deformation,** probably indicates the presence of a particularly weak fault zone within a plate, capable of moving in response to the ambient stress state within a plate. Plate-interior deformation currently occurs along the New Madrid Fault System in the Mississippi Valley of the central United States.

Geoscientists distinguish three types of plate boundaries (Figure 14.14a–d). (1) At **divergent plate boundaries,** defined by mid-ocean ridges (also called oceanic ridges, because not all occur in the middle of an ocean), two plates move apart as a consequence of seafloor spreading. Thus, these boundaries are also called "spreading centers." The process of seafloor spreading produces new oceanic lithosphere (see Chapter 16). (2) At **convergent plate boundaries,** one oceanic plate sinks into the mantle beneath an overriding plate, which can be either a continental or an oceanic plate (see Chapter 17). During this process, which is called **subduction,** an existing oceanic plate gradually disappears. Thus, convergent plate boundaries are also called "destructive boundaries" or "consuming boundaries." Volatile elements (water and carbon dioxide) released from the subducted plate trigger melting in the overlying asthenosphere. The resulting magma rises and erupts in a chain of volcanoes, called a **volcanic arc,** along the edge of the overriding plate. "Arcs" are so named because many, though not all, are curved in map view. The actual plate boundary at a convergent boundary is delineated by a deep ocean **trench.** Convergence direction is not necessarily perpendicular to the trench. Where convergence occurs at an angle of less than 90° to the trend of the trench, the movement is called **oblique convergence.** (3) At **transform plate boundaries,** one plate slides past another along a strike-slip fault. Since no new plate is created and no old plate is consumed along a transform, such a boundary can also be called a "conservative boundary." Transform plate boundaries can occur either in continental or oceanic lithosphere. At some transform boundaries, there is a slight component of convergence, leading to compressive stress. Such boundaries are called **transpressional boundaries.** Similarly, at some transform boundaries there is a slight component of extension, leading to tensile stress. Such boundaries are called **transtensional boundaries.**

In addition to the plate boundaries just discussed, geologists recognize two other locations where movement of lithosphere creates structures. (1) At **collision zones,** two buoyant blocks of crust converge (Figure 14.15a). The buoyant blocks, which may consist of continental crust, island arcs, or oceanic plateaus, are too buoyant to be subducted, so collisions result in broad belts of deformation, metamorphism, and crustal

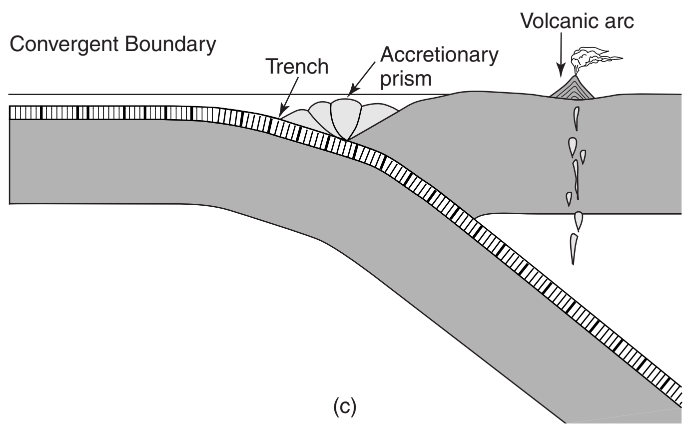


(a)

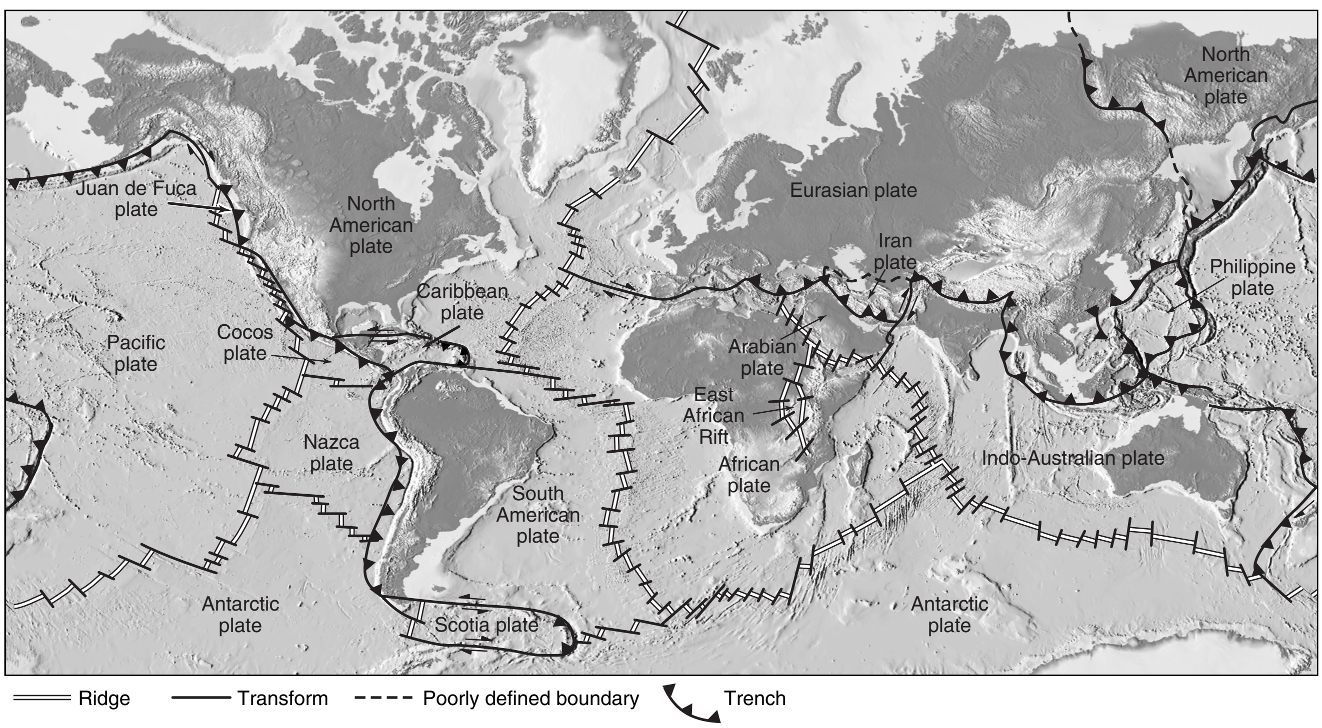


(b)

FIGURE 14.14 The nature of plate boundaries. (a) The basic kinds of plate boundaries. (b) Cross section of a divergent boundary. (c) Cross section of a convergent boundary. (d) The seven major and several minor plates of the Earth. Ocean ridges (double lines), trenches (lines with teeth on the overriding plate), and transform faults (single lines) mark the plate boundaries. In a few places, plate boundaries are ill-defined. These boundaries are marked by dashed lines.



(c)



(d)

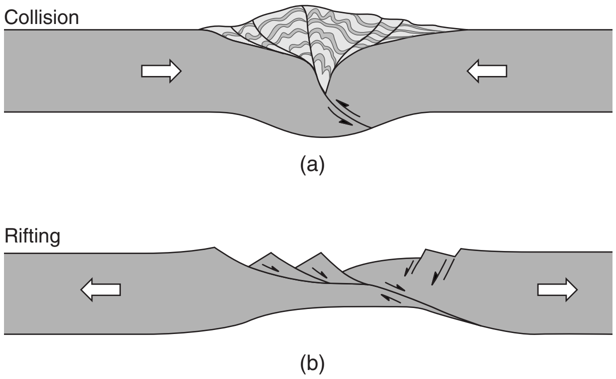


FIGURE 14.15 (a) Schematic cross section of a collision zone, where two buoyant continents converge, creating a broad belt of deformation and crustal thickening. (b) Schematic cross section of a rift, where stretching of the crust causes thinning and normal faulting.

thickening (see Chapter 17). At the end of a collision, two continents that were once separate have become stuck together to form one continuous continent; the boundary here is called a **suture.** Large continents formed when several smaller continents have sutured together are called **supercontinents.** (2) At a **rift,** an existing continent stretches and starts to split apart (Figure 14.15b). At a **successful rift,** the continent splits in two and a new mid-ocean ridge forms. The stretched continental crust along the boundary of a successful rift evolves into a passive continental margin. At an **unsuccessful rift,** rifting stops before the split is complete so the rift remains as a permanent scar in the crust. It is usually marked by an elongate trough, bordered by normal faults and filled with thick sediment and/or volcanic rock.

As shown in Figure 14.13, the Earth currently has seven major plates (Pacific, North American, South American, Eurasian, African, Indo-Australian, and Antarctic Plates) and seven smaller plates (e.g., Cocos, Scotia, and Nazca Plates). In addition, Earth has several microplates. A plate can consist entirely of oceanic lithosphere (such as the Nazca Plate), but most plates consist of both oceanic and continental lithosphere (Figure 14.15). For example, the North American Plate consists of the continent of North America *and* the western half of the Atlantic Ocean. Thus, not all of today's continental margins are plate boundaries, and for this reason, we make the distinction between **active continental margins,** which are plate boundaries, and **passive continental margins,** which are not plate boundaries. The western margin of Africa is a passive margin, while the western margin of South America is an active margin. As noted earlier, passive margins form from the stretched continental crust left by the rifting that led to the successful formation of a new mid-ocean ridge.

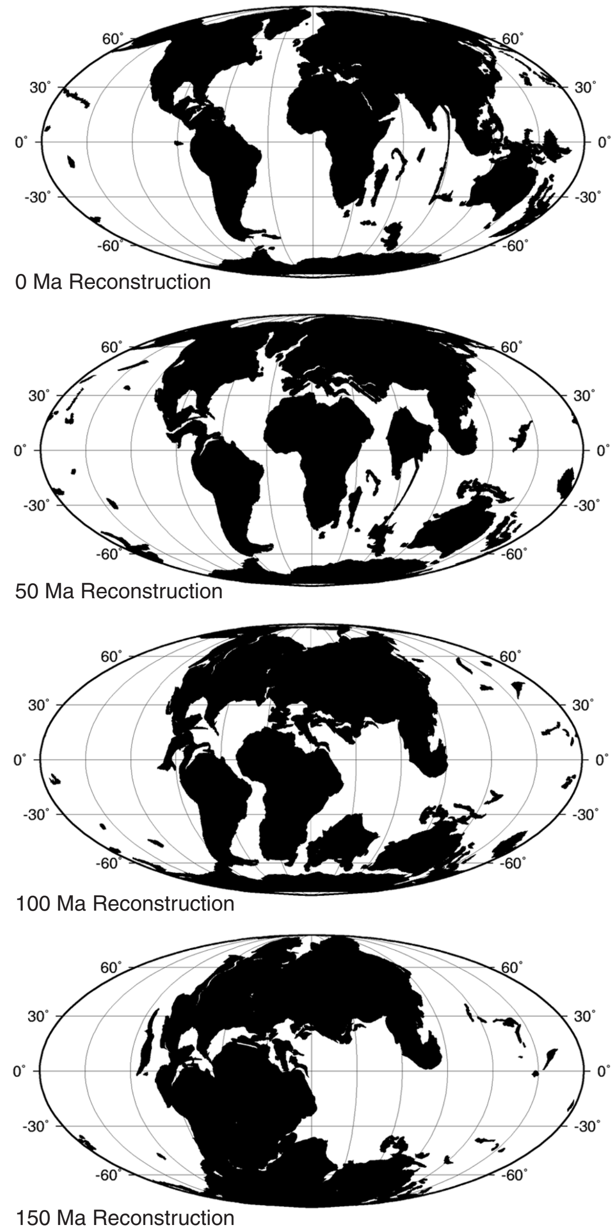


FIGURE 14.16 A sequence of maps showing the motion of continental elements on Earth, following the Early Mesozoic breakup of Pangaea (Molweide projection).

Continents on either side of a mid-ocean ridge move apart as seafloor spreading causes the intervening ocean basin to grow. Continents separated from each other by a subduction zone passively move together as subduction consumes intervening seafloor. Thus, plate motion causes the map of the Earth's surface to constantly change through time, so the process of **continental drift,** as envisioned by Alfred Wegener, does indeed occur (Figure 14.16).

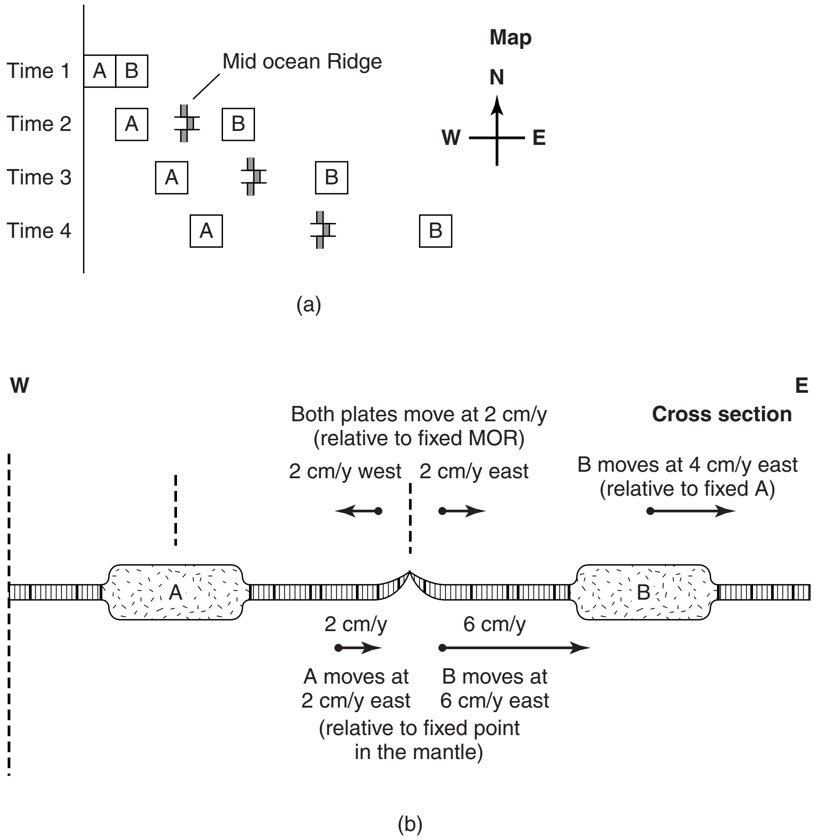


FIGURE 14.17 Describing plate velocity. (a) A sequence of map-view drawings illustrating the movement of two continents on the surface of the Earth, as seafloor spreading occurs between them. Both continents are drifting east, in an absolute reference frame (the line on the left). At the same time, B is moving to the east relative to A, which is equivalent to saying that A is moving to the west relative to B. (b) This cross section illustrates the velocities of plates A and B in different reference frames. First we consider relative velocities. Plate A is moving west at 2 cm/y relative to the mid-ocean ridge (MOR), and plate B is moving east at 2 cm/y relative to the MOR. Thus, plate B moves east at 4 cm/y relative to a fixed plate A, and plate A moves at 4 cm/y west relative to a fixed plate A . Now we consider absolute velocities (i.e., velocities relative to a fixed point in the mantle). Plate A moves east at 2 cm/y relative to a fixed point in the mantle, while plate B moves east at 6 cm/y relative to a fixed point in the mantle. Note that the relative motion of plate B with respect to plate A occurs because plate B's absolute velocity to the east is faster than that of plate A.

14.9 BASIC PLATE KINEMATICS

When we talk of **plate kinematics,** we are referring to a description of the rates and directions of plate motion on the surface of the Earth. Description of plate motion is essentially a geometric exercise, made somewhat complex because the motion takes place on the surface of a sphere. Thus, the description must use the tools of **spherical geometry,** and to do this, we must make three assumptions: First, we must assume that the Earth is, indeed, a sphere. In reality, the Earth is slightly flattened at the poles (the Earth's radius at the poles is about 20 km less than it is at the equator), but this flattening is not sufficiently large for us to worry about. Second, we must assume that the circumference of the Earth remains constant through time. This assumption implies that the Earth neither expands nor contracts, and thus that the amount of seafloor spreading worldwide is equivalent to the amount of subduction worldwide. Note that this assumption does *not* require that the rate of subduction on one side of a specific plate be the same as the rate of spreading on the other side of that plate, but only that growth and consumption of plates are balanced for all plates combined over the whole surface of the Earth. Third, we must assume that plates are internally rigid, meaning that all motion takes place at plate boundaries. This assumption, as we noted earlier, is not completely valid because continental plates do deform internally, but the error in plate motion calculations resulting from this assumption is less than a few percent.

Geoscientists use two different reference frames to describe plate motion (Figure 14.17). When using the **absolute reference frame,** we describe plate motions with respect to a fixed point in the Earth's interior. When using the **relative reference frame,** we describe the motion of one plate with respect to another. To understand this distinction, imagine that you are describing the motion of two cars cruising down the highway. If we say that Car A travels at 60 km/h, and Car B travels at 40 km/h, we are specifying the *absolute* velocity of the cars relative to a fixed point on the ground. However, if we say that Car A travels 20 km/h faster than Car B, we are specifying the *relative* velocity of Car A with respect to Car B. We will next look at how we specify absolute plate velocity and then we will examine relative plate velocity.

14.9.1 Absolute Plate Velocity

We mentioned earlier that not all volcanoes occur along plate boundaries. Some, called hot-spot volcanoes, erupt independently of plate-boundary activity;

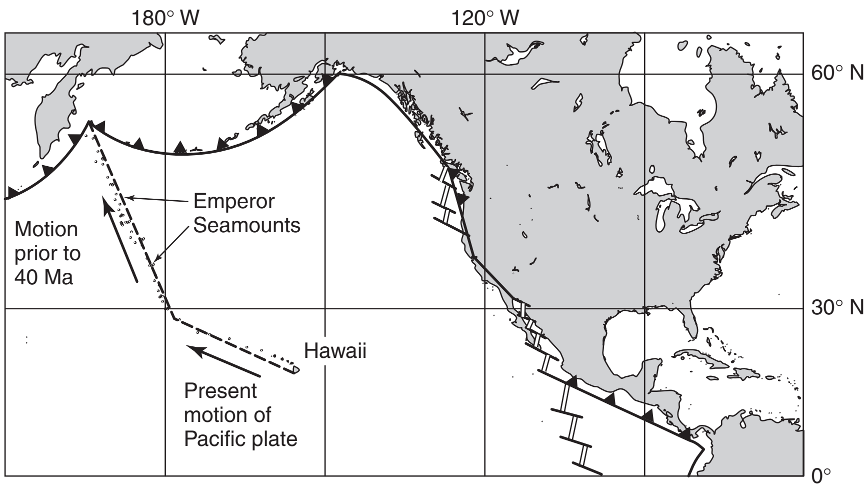


FIGURE 14.18 The Hawaiian-Emperor hot-spot track. The bend shows that the movement of the Pacific Plate relative to the Hawaiian plume changed abruptly at 40 Ma.

they form above a mantle plume. Since mantle plumes appear to be independent of plate boundaries, they can be used as the "fixed" reference points for calculating absolute plate velocities.

To see how this works, picture a lithosphere plate moving over a plume. At a given time, a hot-spot volcano forms above the plume. But as the plate moves, it carries the volcano off of the plume, and when this happens, the volcano stops erupting. Eventually, a new volcano forms above the plume, but as more time passes this volcano will also be carried off the plume. Thus, over time, a chain of volcanic islands and seamounts develops on the lithosphere. As long as the plume exists, one end of this chain will be an active volcano. The age of the volcanoes in the chain increases progressively along the chain away from the plume. These chains are called **hot-spot tracks.** The Hawaiian-Emperor chain in the Pacific Ocean serves as a classic example of a hot-spot track (Figure 14.18). Some plumes are long-lived, lasting 100 million years or more, whereas others are short-lived, lasting less than 10 million years. The orientation of a hotspot track gives the direction of plate motion, and the rate of change in the age of volcanic rocks along the length of the track represents the velocity of the plate.

Note that a distinct bend in a hot-spot track indicates that there has been a sudden change in the direction of absolute motion. For example, the bend in the Hawaiian-Emperor chain reveals that the Pacific Plate moved north-northwest before 40 Ma, but has moved northwest since 40 Ma. Geologists determined the age of this shift by radiometrically dating volcanic rocks from Midway Island, the extinct volcanic island that occurs at the bend.

The hot-spot frame of reference gives a reasonable *approximation* of absolute plate velocity, but, in reality, it's not perfect because hot spots actually do move with respect to one another. Nevertheless, the velocity of hot-spot movement is an order of magnitude less than the velocity of plate movement, so the error in absolute plate motions based on the hot-spot reference frame is only a few percent.[8]

Figure 14.19 shows the absolute velocity of plates. You will notice that absolute plate rates today range from less than 1 cm/y (for the Antarctic and African Plates) to about 10 cm/y (for the Pacific Plate). North America and South America are moving west at about 2–3 cm/y. For comparison, these rates are about the same at which your hair or fingernails grow. This may seem slow, but remember that a velocity of only 2 cm/y can yield a displacement of 2000 km in 100 million years! There has been plenty of time during Earth's history for large oceans to open and close many times.

14.9.2 Relative Plate Velocity

The motion of any plate with respect to another can be defined by imagining that the position of one of the plates is fixed. For example, if we want to describe the motion of plate A with respect to plate B, we fix

[8]To get more accurate measurements of absolute plate motion, geoscientists use the **no net torque calculation.** In this calculation, we assume that the sum of all the plate movements shearing against the underlying asthenosphere creates zero torque on the asthenosphere. The no net torque calculation is well suited for computer calculations. See Cox and Hardt (1986) for details.

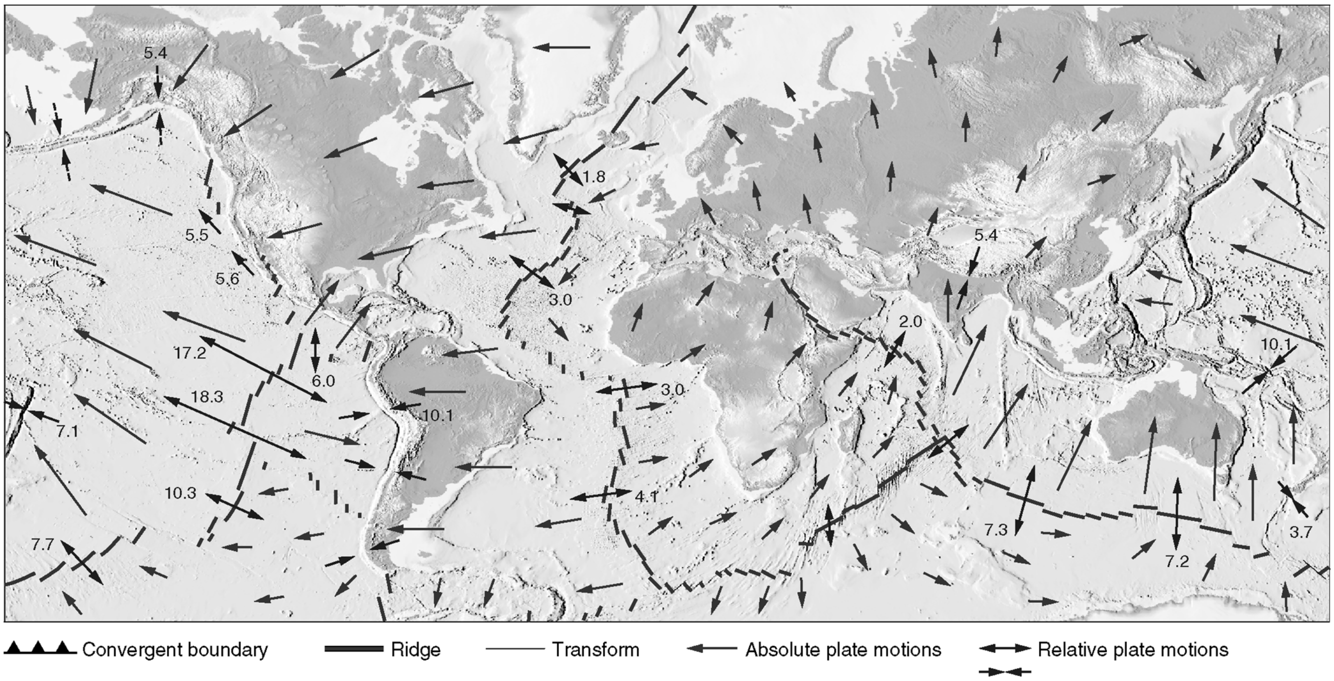


FIGURE 14.19 Directions and rates in mm/y of plate motion. Relative motions of the major plates are given for selected points along their boundaries by pairs of inward-pointing (convergent) or outward-pointing (divergent) arrows. The numbers next to these arrows give the relative velocity in cm/y.

plate B on the surface of the Earth and see how plate A moves (Figure 14.17). Recall that in plate kinematic calculations, the Earth is considered to be a sphere, so such a movement is described by a rotation at a specified angular velocity around an imaginary axis that passes through the center of the Earth. The intersection between this imaginary rotation axis and the surface of the Earth is called an **Euler pole** (Figure 14.20a–c). Note that Euler poles are merely geometric elements and that they are not related to the Earth's geographic poles (the points where the Earth's spin axis intersects the surface), nor are they related to the Earth's magnetic poles (the points where the Earth's internal dipole intersects the surface).

To avoid confusion, it is useful to distinguish between two types of Euler poles. An **instantaneous Euler pole** can be used to describe relative motion between two plates at an instant in geologic time, whereas a **finite Euler pole** can be used to describe total relative motion over a long period of geologic time. For example, the *present-day* instantaneous Euler pole describing the motion of North America with respect to Africa can be used to determine how fast Chicago is moving away from Casablanca today. We could calculate a single finite Euler pole to describe the motion between these two locations over the past 80 million years, even if the instantaneous Euler pole changed at several times during this interval.

14.9.3 Using Vectors to Describe Relative Plate Velocity

Let's now see how to describe relative plate motion with the use of vectors. Imagine two plates, A and B, that are moving with respect to each other. We can define the relative motion of plate A with respect to plate B by the vector ${}_{A}\mathbf{\Omega}_{B}$, where

$$ {}_{A}\mathbf{\Omega}_{B} = \omega\mathbf{k} \qquad \text{Eq. 14.1}$$

In this formula, ω is the angular velocity and $\mathbf{k}$ is a unit vector parallel to the rotation axis.

For most discussions of plate kinematics, however, it is easier for us to describe motion in terms of the **linear velocity, v,** as measured in centimeters per year at a point on a plate. For example, we can say that New York City, a point on the North American Plate, is moving west at 2.5 cm/y with respect to Paris, a point on the Eurasian Plate. Note that we can *only* describe $\mathbf{v}$ if we specify the point at which $\mathbf{v}$ is to be measured. If we know the value of ${}_{A}\mathbf{\Omega}_{B}$ defining the relative plate motion, we can calculate the value of $\mathbf{v}$ at a point.

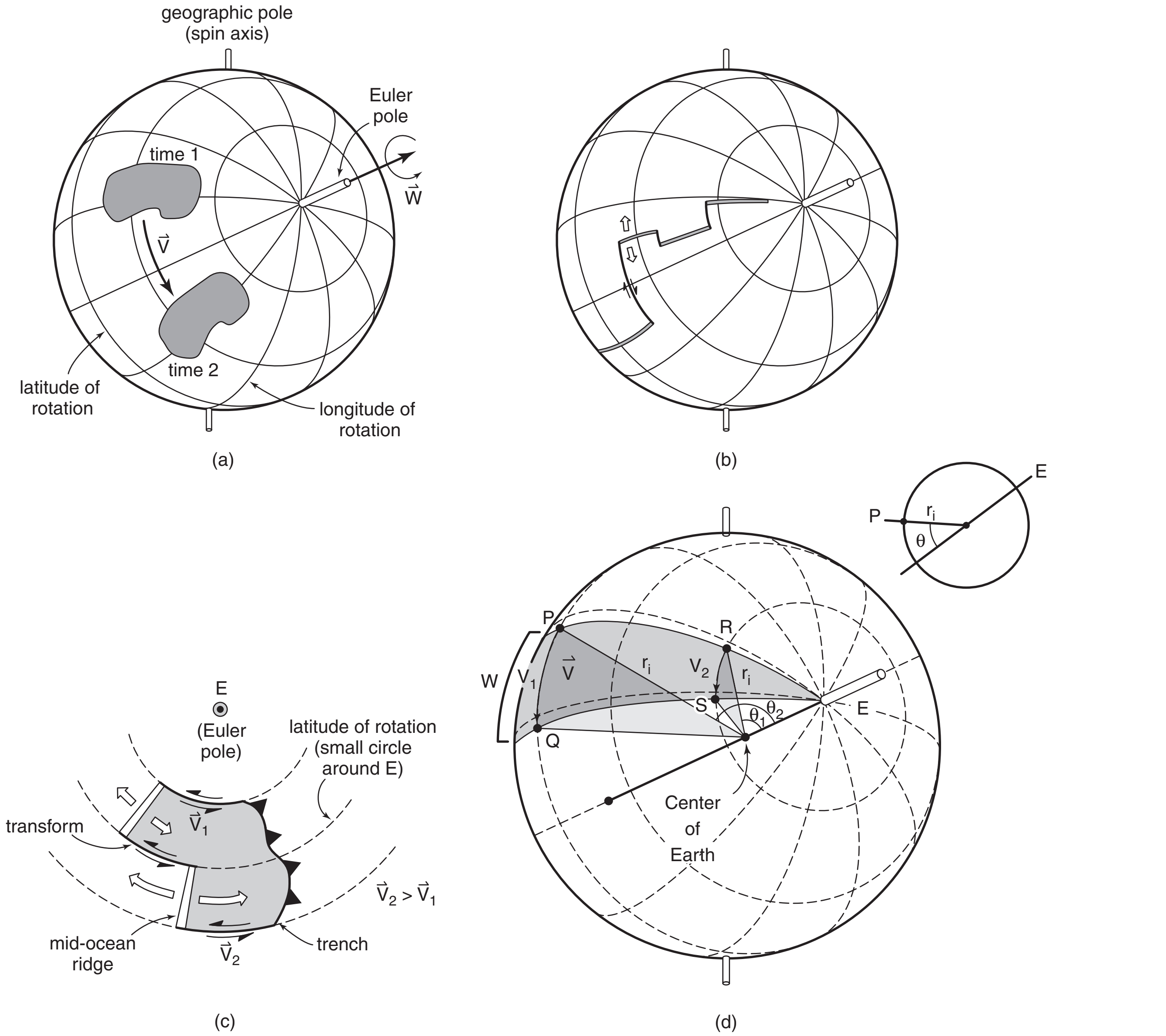


FIGURE 14.20 Describing the motion of plates on the Earth's surface. (a) Latitudes and longitudes around the Euler pole (E) are small circles and great circles, respectively. The motion of a plate between time 1 and time 2 is a **rotation** **(**ω**)** around the Euler axis. The linear velocity **(v)** is measured on the surface of the Earth. (b) Transform faults are small circles (latitudes of rotation) relative to the Euler pole. (c) We can also see the relation of transforms relative to the Euler pole in a projection looking down on the Euler pole. Note that spreading velocity and rate of slip along transform faults increase with increasing distance away from the Euler pole. (d) Linear velocity increases as the angle θ increases. r_i is the radius vector to a point on the Earth's surface. Thus, since $\theta_2 > \theta_1$, $V_2 > V_1$.

Specifically, **v** is the vector cross product of ${}_A\boldsymbol{\Omega}_B$ and the radius vector ($\mathbf{r}_i$) drawn from the center of the Earth to the point in question. This relation can be represented by the equation

$$\mathbf{v} = {}_A\boldsymbol{\Omega}_B \text{ X } \mathbf{r}_i \qquad \text{Eq. 14.2.}$$

As in any vector cross product, this equation can be rewritten as

$$\mathbf{v} = \mathbf{r}_i \sin \theta \qquad \text{Eq. 14.3}$$

where θ is the angle between $\mathbf{r}_i$ and the Euler axis (Figure 14.20d).

We see from the preceding equations that **v** is a function of the distance along the surface of the Earth between the point at which **v** is determined and the Euler pole. As you get closer to the Euler pole, the value of θ becomes progressively smaller, and at the pole itself, $\theta = 0°$. Since $\sin 0° = 0$, the relative linear velocity (**v**) between two plates *at* the Euler pole is 0 cm/y. To picture this relation, think of a Beatles record playing on a stereo. Even when the record spins at a constant angular velocity, the linear velocity at the center of the record is zero and increases toward the edge of the record. Thinking again of plate motions, note that the maximum relative linear velocity occurs where $\sin \theta = 1$ (i.e., at 90° from the Euler pole). What this means is that the relative linear velocity between two plates changes along the length of a plate boundary. For example, if the boundary is a mid-ocean ridge, the spreading rate is greater at a point on the ridge at 90° from the Euler pole than it is at a point close to the Euler pole. Note that in some cases, the Euler pole lies on the plate boundary, but it does not have to be on the boundary. In fact, for many plates the Euler pole lies off the plate boundary.

Because the relative velocity between two plates can be described by a vector, plate velocity calculations obey the **closure relationship,** namely

$$_{A}\mathbf{\Omega}_{C} = {_{A}\mathbf{\Omega}_{B}} + {_{B}\mathbf{\Omega}_{C}} \qquad \text{Eq. 14.4}$$

Using the closure relationship, we can calculate the relative velocity of two plates even if they do not share a common boundary. For example, to calculate the relative motion of the African Plate with respect to the Pacific Plate, we use the equation:

$$_{\text{Africa}}\mathbf{\Omega}_{\text{Pacific}} = {_{\text{Africa}}\mathbf{\Omega}_{\text{S. America}}} + {_{\text{S. America}}\mathbf{\Omega}_{\text{Nasca}}} + {_{\text{Nasca}}\mathbf{\Omega}_{\text{Pacific}}} \qquad \text{Eq. 14.5}$$

An equation like this is called a **vector circuit.**

At this point, you may be asking yourself the question, how do we determine a value for $_{A}\mathbf{\Omega}_{B}$ in the first place? Actually, we can't measure $_{A}\mathbf{\Omega}_{B}$ directly. We must calculate it by knowing ω, which we determine, in turn, from a knowledge of **v** at various locations in the two plates or along their plate boundary. We can measure values for **v** directly along divergent and transform boundaries.

As an example of determining **v** along a divergent boundary, picture point P on the Mid-Atlantic Ridge, the divergent boundary between the African and the North American Plates. To determine the instantaneous Euler pole and the value for **v** at point P, describing the motion between these two plates, we go through the following steps:

- First, since **v** is a vector, we need to specify the *orientation* of **v,** in other words, the spreading direction. To a first approximation, the spreading direction is given by the orientation of the transform faults that connect segments of the ridge, for on a transform fault, plates slip past one another with no divergence or convergence. Geometrically, a transform fault describes a small circle around the Euler pole (Figure 14.20), just as a line of latitude describes a small circle around the Earth's geographic pole. Therefore, the direction of **v** at point P is parallel to the nearest transform fault.
- Second, we need to determine the location of the Euler pole describing the motion of Africa with respect to North America. Considering that transform faults are small circles, a great circle drawn perpendicular to a transform fault must pass through the Euler pole (Figure 14.21), just as geographic lines of longitude (great circles perpendicular to lines of latitude) must pass through the geographic pole. So, to find the position of the Euler pole, we draw great circles perpendicular to a series of transforms along the ridge, and where these great circles intersect is the Euler pole.
- Finally, we need to determine the magnitude of **v.** To do this, we determine the age of the oceanic crust on either side of point P.[9] Since velocity is distance divided by time, we simply measure the distance between two points of known, equal age on either side of the ridge to calculate the spreading velocity across the ridge. This gives us the magnitude of **v** at point P.

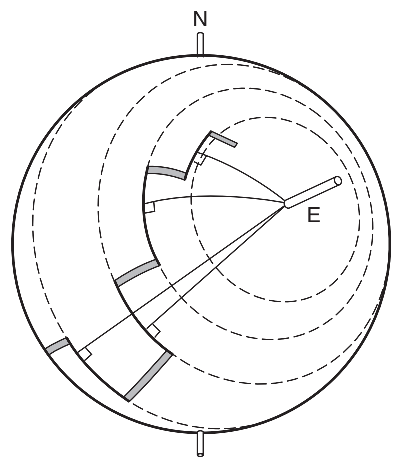


FIGURE 14.21 The location of an Euler pole can be determined by drawing a great circle (longitude of rotation) perpendicular to each of the transforms. The point at which the great circles meet is the Euler pole (E).

On continental transform boundaries, like the San Andreas Fault in California, the magnitude of **v** can be determined by matching up features of known age that

[9]We can determine the age of the crust by dating the fossils directly at the contact between the pillow basalt layer and the pelagic sediment layer, or we can use the age of known "marine magnetic anomalies." We will discuss these anomalies in Chapter 16.

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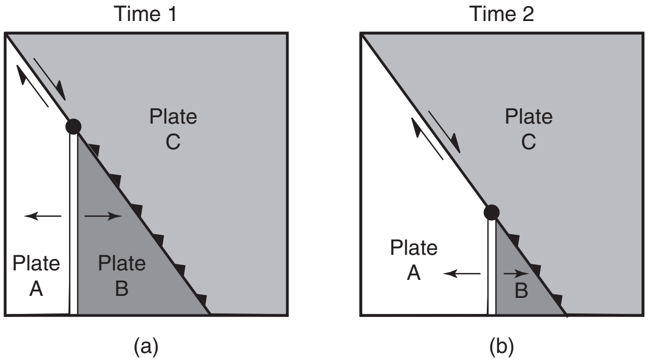

FIGURE 14.22 Geometry of triple junctions. (a) This ridge-trench-transform triple junction is stable. (b) With time the ridge-trench-transform triple junction location changes, but the geometry stays the same.

formed across the fault after the fault formed, but are now offset by the fault. We cannot directly determine relative motion across a convergent boundary, so the value of **v** across convergent boundaries must be determined using vector circuits.

Until the end of the twentieth century, relative plate motion could not be confirmed by direct observation. But now, the **global positioning system** (GPS), which uses signals from an array of satellites to determine the location of a point on Earth's surface, makes this possible. By setting up a network of GPS stations, it is possible to define the location of a point to within a few millimeters, and with this accuracy, plate movements over a period of a few months to years can indeed be detected.

14.9.4 Triple Junctions

At this point, you should have a fairly clear idea of what we mean by a "plate boundary" between two plates. We can represent the intersection of a plate boundary with the surface of the Earth by a line on a map. The point where three plate boundaries meet is called a **triple junction.** You can see several examples of triple junctions in Figure 14.14d. We name specific types of triple junction by listing the plate boundaries that intersect. For example, at a ridge-ridge-ridge triple junction, three divergent boundaries intersect (e.g., in the southern South Atlantic and in the southern Indian Ocean), and at a ridge-trench-transform triple junction, a divergent boundary, a convergent boundary, and a transform boundary intersect (e.g., off the west coast of northern California).

Geoscientists distinguish between stable and unstable triple junctions. The basic configuration of a **stable triple junction** can exist for a long time, even though the location of the triple junction changes. For example, the ridge-trench-transform junction in Figure 14.22a is a stable triple junction, because even though the location of the triple junction (point T) migrates to the southeast with time, the geometry stays the same. In contrast, the basic geometry of an **unstable triple junction** must change rapidly to create a new arrangement of plate boundaries.

The migration of a triple junction along a plate boundary can lead to the transformation of a segment of a plate boundary from one type of boundary to another. Such a change occurred along the coast of California, an active margin, during the Cenozoic. Through most of the Mesozoic, and into the Early Cenozoic, the western North American margin was a convergent plate boundary. Toward the end of this time interval, convergence occurred between the North American and Farallon Plates. At around 30 Ma, the Farallon-Pacific Ridge (the divergent boundary between the Farallon and Pacific Plates) began to be subducted. When this occurred, the Pacific Plate came into contact with the North American Plate and two triple junctions formed, one moving north-northwest, and the other moving south-southeast. The margin between the triple junction changed from being a convergent boundary into a transform boundary, the San Andreas Fault.

14.10 PLATE-DRIVING FORCES

Alfred Wegener was unable to convince the geologic community that continental drift happens, because he could not explain *how* it happens. The question of what drives the plates remains controversial to this day. In the years immediately following the proposal of plate tectonics, many geoscientists tacitly accepted a **convection-cell model,** which stated that convection-driven flow in the mantle drives the plates. In this model, plates were carried along on the back of flowing asthenosphere, which was thought to circulate in simple elliptical (in cross section) paths; **upwelling** (upward flow) of hot asthenosphere presumably occurred at mid-ocean ridges, while **downwelling** (downward flow) of hot asthenosphere occurred at the margins of oceans or at subduction zones. In this model, the flowing asthenosphere exerts **basal drag,** a shear stress, on the base of the plate, which is sufficient to move the plate. This image of plate motion, however, was eventually discarded for, while it is clear that the mantle does

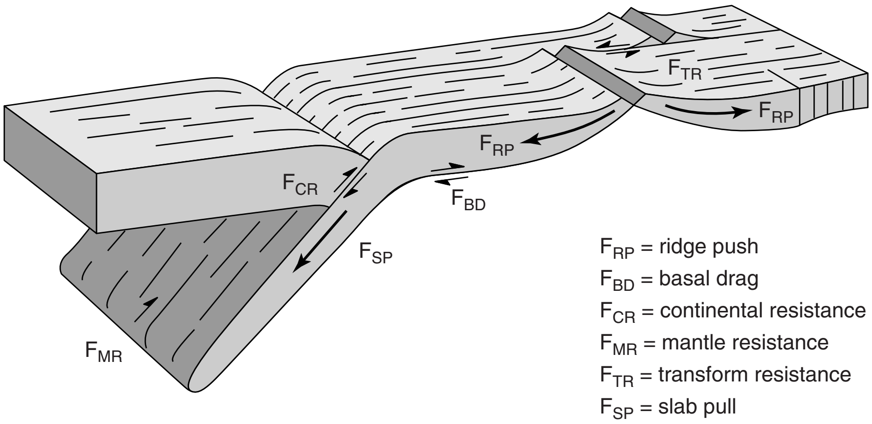

FIGURE 14.23 Simplified diagram illustrating the forces acting on a plate.

convect,[10] it is impossible to devise a global geometry of convection cells that can explain the observed geometry of plate boundaries that now exist on Earth. Subsequent calculations showed that two other forces, ridge push and slab pull, play a major role in driving plates.

Ridge-push force is the outward-directed force that pushes plates away from the axis of a mid-ocean ridge. It exists because oceanic lithosphere is higher along mid-ocean ridges than it is in the abyssal plains at a distance away from the ridge. In Earth's gravity field, the difference in elevation means that the lithosphere along the ridge has more gravitational potential energy than the lithosphere of the abyssal plain, and this energy provides an outward push perpendicular to the ridge axis. To picture this push, imagine a thin pool of molasses on a table top. If you pour more molasses onto the center of the sheet, a mound of molasses builds up in the center. The new molasses flows outward and pushes the molasses that was already on the table in front if it, causing the diameter of the molasses sheet to increase.

Slab-pull force is the force that pulls lithosphere into a convergent margin. It exists because old, cold ocean lithosphere is negatively buoyant relative to the underlying asthenosphere, so if given a chance, the oceanic lithosphere sinks downwards. Once the cold subducting plate (also called a subducted "slab") descends into the mantle, phase transformations change basalt to much denser eclogite, so a subducted plate is even denser than a plate at the Earth's surface. Thus, like a sinking anchor, a subducted slab "pulls" the rest of the plate down with it. To picture the process, imaging that you smooth a sheet of wet tissue paper onto the ceiling. If you peel one end of the paper off, and let go, the paper slowly moves down and peels the rest of the paper off with it, until the whole sheet eventually separates from the ceiling and falls.

Let's recap and consider all the forces that act on plates (Figure 14.23). Ridge-push forces drive plates away from mid-ocean ridges, and slab-pull forces drive them down subduction zones. Thus, plate motion is, to some extent, a *passive* phenomenon, in that it is the gravitational potential energy of the plate itself that causes a plate to move. Does basal drag due to mantle convection play a role at all in driving plates? Probably yes, but not in the simple way that researchers first envisioned. Convection in the mantle, according to seismic tomography studies, seems to be accommodated *globally* by a few upwelling zones and a few downwelling zones. These zones do not exactly correspond to the present configuration of ridges and trenches. The flow of asthenosphere due to convection probably does create some basal drag at the base of plates, but the basal drag force can assist *or* retard motion. Specifically, where the asthenosphere flows in the same direction as the plate motion caused by ridge push and/or slab pull, basal drag may accelerate the motion. In contrast, where the asthenosphere flows in a direction opposite to the plate motion caused by ridge push and/or slab pull, it may retard the motion. And if the asthenosphere flows at an angle to the plate motion caused by ridge push and/or slab pull, it may change the direction of motion. Calculations show that the size of the basal-drag force caused by asthenosphere flow is less than that of ridge push or slab pull, and thus basal drag is not the dominant force. The forces that drive plates are resisted by frictional forces between plates.

Note that even though circulation of the asthenosphere alone does not drive plate motion, the mobility of the asthenosphere does make plate motion possible. If the asthenosphere could not flow up at mid-ocean ridges to fill space created by seafloor spreading, and if the asthenosphere could not move out of the way of subducting slabs, then buoyancy forces (negative or positive) would not be sufficient to cause plates to move. Further, the formation of new hot lithosphere at mid-ocean ridges and the eventual sinking of this lithosphere, once it has cooled, at trenches is itself a form of convection.

[10] Seismic tomography shows variations in temperature in the mantle, and submarine measurements show that heat flow is higher at mid-ocean ridges than elsewhere.

14.11 THE SUPERCONTINENT CYCLE

While Wegener's view that continental drift happens ultimately proved correct, his view that drift *only* happened after the breakup of Pangaea was just part of the story. J. Tuzo Wilson, a Canadian geophysicist, noted that formation of the Appalachian-Caledonide Orogen involved closure of an ocean. Thus, there must have been another ocean (not the Atlantic) on the east side of North America *before* the formation of Pangaea. Wilson envisioned a cycle of tectonic activity in which an ocean basin opens by rifting, grows by seafloor spreading, closes by subduction, and disappears during collision and supercontinent formation and, further, that more than one such cycle has happened during Earth history. The successive stages of rifting, seafloor spreading, convergence, collision, rifting, recorded in a single mountain range came to be called the **Wilson cycle.** Because of the Wilson cycle, we cannot find oceanic lithosphere older than about 200 Ma on Earth—it has all been subducted. Very old continental crust, however, can remain because it's too buoyant to be subducted; that's why we can still find Archean rocks on continents.

Eventually, geologists realized that Wilson cycles were part of a global succession of events leading to formation of a supercontinent, breakup of a supercontinent, dispersal of continents, and reassembly of continents into a new supercontinent. This succession of events has come to be known as the **supercontinent cycle** (Figure 14.24). At various times in the past, continental movements and collisions have produced supercontinents, which lasted for tens of millions of years before eventually rifting apart. The geologic record shows that a supercontinent, Pangaea, had formed by the end of the Paleozoic (~250 Ma), only to disperse in the Mesozoic. Similarly, a supercontinent formed at the end of the Precambrian (1.1 Ga, called Rodinia), only to disperse at about 900 Ma to form a new supercontinent, Panotia, which itself broke up about 600 Ma (i.e., at the beginning of the Paleozoic). And there is growing evidence that supercontinents also formed even earlier in Earth history. Thus, it seems that supercontinents have formed, broken up, formed, and broken up at roughly 200–500 m.y. intervals.

Modeling suggests that the supercontinent cycle may be related to long-term convection patterns in the mantle. In these models, relative motion between plates at any given time—the "details" of plate kinematics—is determined by ridge-push and slab-pull forces. But over long periods of time (about 200–500 m.y.), continents tend to accumulate over a zone of major mantle

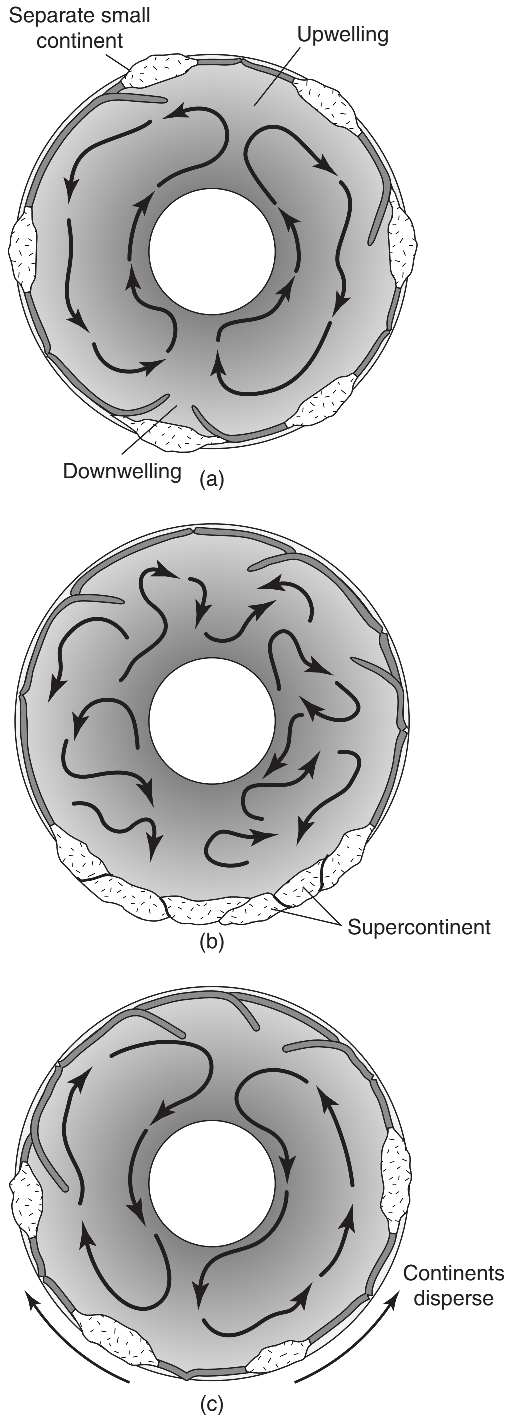

FIGURE 14.24 Stages in the supercontinent cycle. (a) Continents move relative to one another, but gradually aggregate over a mantle downwelling zone. (b) While the supercontinent exists, large-scale convection in the mantle reorganizes. (c) Upwelling begins beneath the supercontinent and weakens it. Eventually, rifting occurs and the supercontinent breaks up, forming smaller continents that drift apart.

downwelling to form a supercontinent (Figure 14.18). However, supercontinents don't last forever; once a supercontinent forms, the thermal structure of the mantle beneath it changes. This change happens because a supercontinent acts like a giant insulator that does not allow heat to escape from the mantle. Over 80% of the Earth's internal heat escapes at mid-ocean ridges, where seawater circulating through the hot crust and upper mantle cools the lithosphere much like a coolant cools an automobile engine; however, within the area of a supercontinent, there are no ridges, so heat cannot easily be lost. As a consequence, the mantle beneath the supercontinent eventually heats up, and can no longer be a region of downwelling. When this happens, a new downwelling zone develops elsewhere, and hot asthenosphere must begin to upwell beneath the supercontinent. Upwelling causes the continental lithosphere of the supercontinent to heat up and weaken. Ultimately, the supercontinent rifts apart into smaller continents separated by new oceans.

If the supercontinent-cycle model works, then we can say that Earth is currently in the dispersal stage of the cycle. During the next 200 million years, the assembly stage will begin and the floors of the Atlantic and Pacific Oceans will be subducted. After a series of collisions, the continents will once again coalesce to form a new supercontinent.[11]

14.12 CLOSING REMARKS

A complete description of whole-Earth structure and plate-tectonics theory would each fill an entire book; in fact, many such books have been written (see Additional Reading). Here, our purpose was simply to remind you of some of the key features of whole-Earth structure, and some of the key tenets of the theory. This information sets the stage for our discussion of structures and other geologic features that develop in specific kinds of tectonic settings. After reviewing basic geophysical methods, we introduce extensional tectonics, meaning the geology of rifts and passive margins. You will see that understanding the architecture of rifts and passive margins is a prerequisite for understanding the consequences of collision and convergence. You will find that many of the topics discussed in this and subsequent chapters will resurface in the essays on representative orogens in Chapter 22.

[11]The hypothetical supercontinent formed by the predicted collision of the Americas and Asia has been called "Amasia."

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