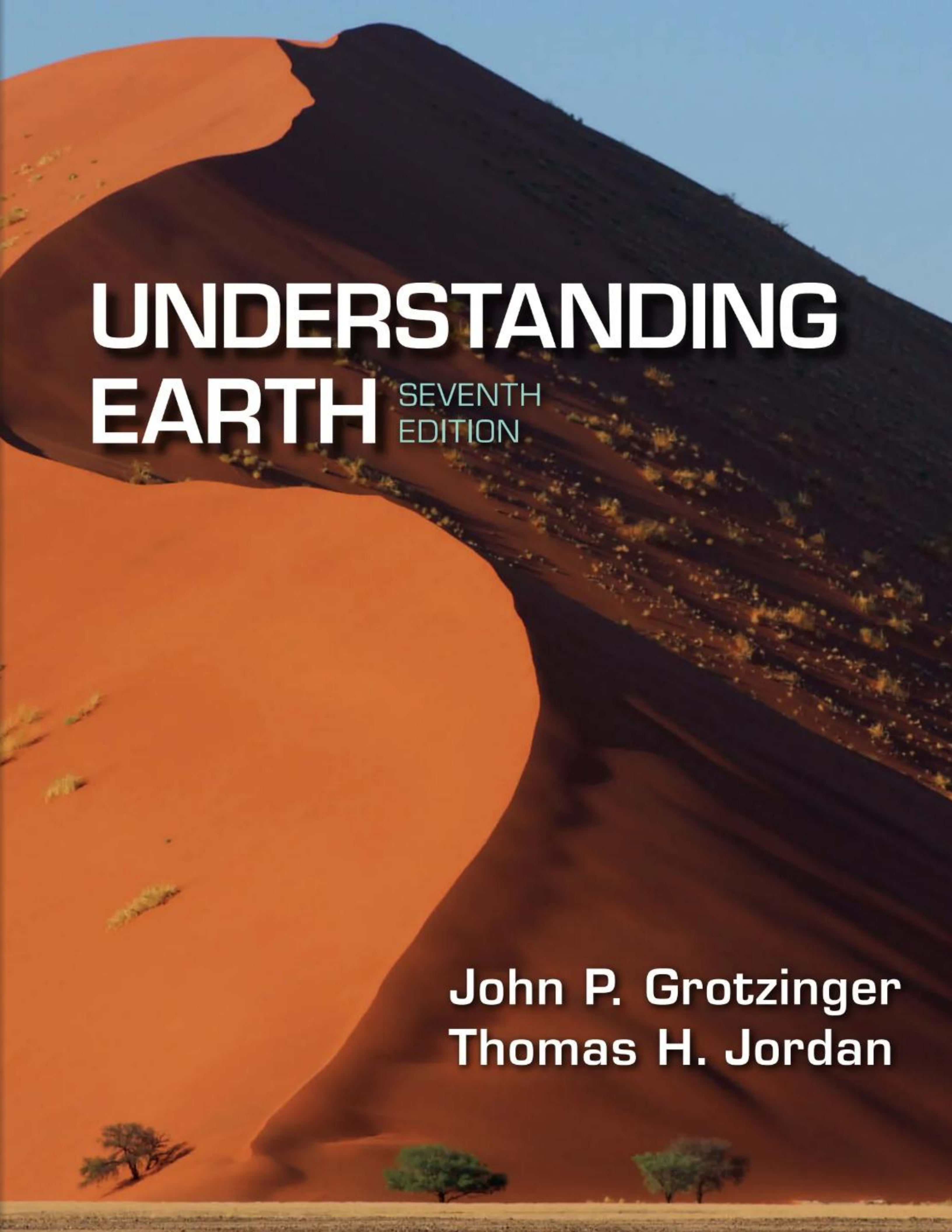


UNDERSTANDING EARTH

SEVENTH EDITION

John P. Grotzinger
Thomas H. Jordan

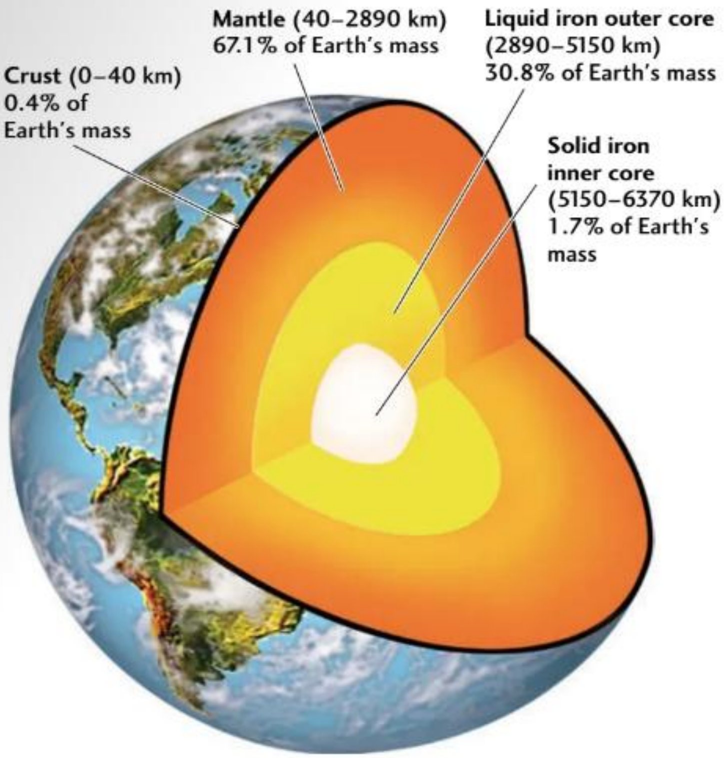


FIGURE 1.9 ■ Earth's major layers, showing their depths and their masses expressed as a percentage of Earth's total mass.

Earth's interior was divided into concentric layers of different compositions, separated by sharp, nearly spherical boundaries (**Figure 1.9**).

Earth's Density

Layering of Earth's deep interior was first proposed by the German physicist Emil Wiechert at the end of the nineteenth century, before much seismic data had become available. He wanted to understand why our planet is so heavy, or more precisely, so *dense*. The density of a substance is easy to calculate: just measure its mass on a scale and divide by its volume. A typical rock, such as the granite used for tombstones, has a density of about 2.7 grams per cubic centimeter (g/cm^3). Estimating the density of the entire planet is a little harder, but not much. Eratosthenes had shown how to measure Earth's volume in 250 B.C., and sometime around 1680, the great English scientist Isaac Newton figured out how to calculate its mass from the gravitational force that pulls objects to its surface. The details, which involved careful laboratory experiments to calibrate Newton's law of gravity, were worked out by another Englishman, Henry Cavendish. In 1798, he calculated Earth's average density to be about 5.5 g/cm^3, twice that of tombstone granite.

Wiechert was puzzled. He knew that a planet made entirely of common rocks could not have such a high density. Most common rocks, such as granite, contain a high proportion of silica (silicon plus oxygen; SiO_2) and have relatively low densities, below 3 g/cm^3. Some iron-rich rocks brought to Earth's surface by volcanoes have densities as high as 3.5 g/cm^3, but no ordinary rock approached Cavendish's value. He also knew that, going downward into Earth's interior, the pressure on rock increases with the weight of the overlying mass. The pressure squeezes the rock into a smaller volume, making its density higher. But Wiechert found that even the effect of pressure was too small to account for the density Cavendish had calculated.

The Mantle and Core

In thinking about what lay beneath his feet, Wiechert turned outward to the solar system and, in particular, to meteorites, which are pieces of the solar system that have fallen to Earth. He knew that some meteorites are made of an *alloy* (a mixture) of two heavy metals, iron and nickel, and thus have densities as high as 8 g/cm^3 (**Figure 1.10**). He also knew that these two elements are relatively abundant throughout our solar system. So, in 1896, he proposed a grand hypothesis: sometime in Earth's past, most of the iron and nickel in its interior had dropped inward to its center under the force of gravity. This movement created a dense **core,** which was surrounded by a shell of silicate-rich rock that he called the **mantle** (using the German word for "coat"). With this hypothesis, he could come up with a two-layered Earth model that agreed with Cavendish's value for Earth's average density. He could also explain the existence of iron-nickel meteorites: they were chunks from the core of an Earthlike planet (or planets) that had broken apart, most likely by collision with other planets.

Wiechert got busy testing his hypothesis using seismic waves recorded by seismographs located around the globe (he designed one himself). The first results showed a shadowy inner mass that he took to be the core, but he had problems identifying some of the seismic waves. These waves come in two basic types: *compressional waves,* which expand and compress the material they move through as they travel through a solid, liquid, or gas; and *shear waves,* which move the material from side to side. Shear waves can propagate only through solids, which resist shearing, and not through fluids (liquids or gases) such as air and water, which have no resistance to this type of motion.

In 1906, a British seismologist, Robert Oldham, was able to sort out the paths traveled by these two types of seismic waves and show that shear waves did not propagate through the core. The core, at least in its outer part, was liquid! This finding turns out to be not too surprising. Iron melts at a lower temperature than silicates, which is why metallurgists can use containers made of ceramics (which are silicate materials) to hold molten iron. Earth's deep interior is hot enough to melt an iron-nickel alloy, but not silicate rock. Beno Gutenberg, one of Wiechert's students, confirmed Oldham's observations and, in 1914, determined that the depth of the *core-mantle boundary* was about 2890 km (see Figure 1.9).

(a)

(b)

FIGURE 1.10 ■ Two common types of meteorites. (a) This stony meteorite, which is similar in composition to Earth's silicate mantle, has a density of about 3 g/cm^3. (b) This iron-nickel meteorite, which is similar in composition to Earth's core, has a density of about 8 g/cm^3. [John Grotzinger/Ramón Rivera-Moret/Harvard Mineralogical Museum.]

The Crust

Five years earlier, a Croatian scientist had detected another boundary at the relatively shallow depth of 40 km beneath the European continent. This boundary, named the *Mohorovičić discontinuity* (Moho for short) after its discoverer, separates a **crust** composed of low-density silicates, which are rich in aluminum and potassium, from the higher-density silicates of the mantle, which contain more magnesium and iron.

Like the core-mantle boundary, the Moho is a global feature. However, it was found to be substantially shallower beneath the oceans than beneath the continents. On average, the thickness of oceanic crust is only about 7 km, compared with almost 40 km for continental crust. Moreover, rocks in the oceanic crust contain more iron, and are therefore denser, than continental rocks. Because continental crust is thicker but less dense than oceanic crust, the continents ride higher by floating like buoyant rafts on the denser mantle (**Figure 1.11**), much as icebergs float on the ocean. Continental buoyancy explains the most striking feature of Earth's surface topography: why the elevations shown in Figure 1.8 fall into two main groups, 0 to 1 km above sea level for much of the land surface and 4 to 5 km below sea level for much of the deep sea.

Shear waves travel well through the mantle and crust, so we know that both are solid rock. How can continents float on solid rock? Rock can be solid and strong over the short term (seconds to years), but weak over the long term (thousands to millions of years). The mantle below a depth of about 100 km has little strength, and over very long periods, it flows as it adjusts to support the weight of continents and mountains.

The Inner Core

Because the mantle is solid and the outer part of the core is liquid, the core-mantle boundary reflects seismic waves, just as a mirror reflects light waves. In 1936, Danish seismologist

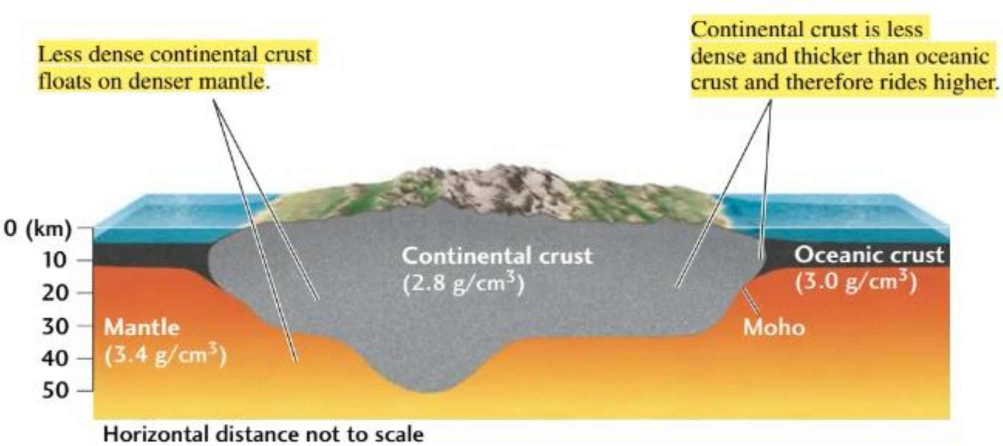


FIGURE 1.11 ■ Because crustal rocks are less dense than mantle rocks, Earth's crust floats on the mantle. Continental crust is thicker and has a lower density than oceanic crust, which causes it to ride higher, explaining the difference in elevation between continents and the deep seafloor.

Inge Lehmann discovered another sharp spherical boundary at the much greater depth of 5150 km, indicating a central mass with a higher density than the liquid core. Studies following her pioneering research showed that the inner core can transmit both shear waves and compressional waves. The **inner core** is therefore a solid metallic sphere suspended within the liquid **outer core**—a "planet within a planet." The radius of the inner core is 1220 km, about two-thirds the size of the Moon.

Geologists were puzzled by the existence of this "frozen" inner core. They knew that temperatures inside Earth should increase with depth. According to the best current estimates, Earth's temperature rises from about 3500°C at the core-mantle boundary to almost 5000°C at its center. If the inner core is hotter, how could it be solid while the outer core is molten? The mystery was eventually solved by laboratory experiments on iron-nickel alloys, which showed that the "freezing" was due to higher pressures, rather than lower temperatures, at Earth's center.

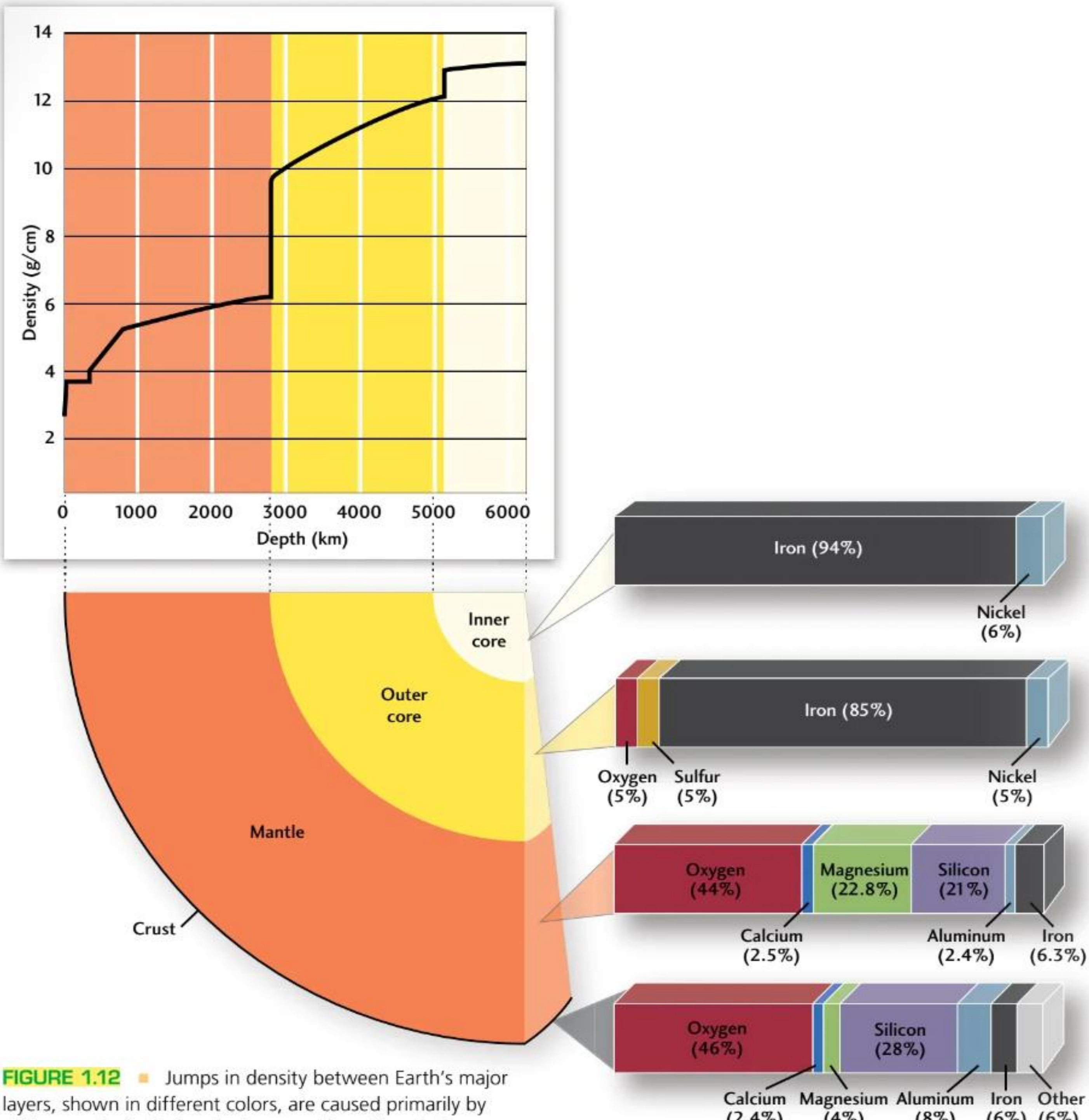


FIGURE 1.12 ■ Jumps in density between Earth's major layers, shown in different colors, are caused primarily by differences in their chemical composition. The relative amounts of the main elements are depicted in the bars on the right.

The Plate Tectonic System

Some of Earth's most dramatic geologic events—volcanic eruptions and earthquakes, for example—result from interactions within Earth's interior. These phenomena are driven by Earth's internal heat, which is transferred upward through the circulation of material in Earth's mantle.

We have seen that Earth is zoned by chemistry: its crust, mantle, and core are chemically distinct layers. Earth is also zoned by *strength*, a property that measures how much an Earth material can resist being deformed. Material strength depends on both chemical composition (bricks are strong, soap bars are weak) and temperature (cold wax is strong, hot wax is weak).

In some ways, the outer part of the solid Earth behaves like a ball of hot wax. Cooling of the surface forms a strong outer shell, or **lithosphere** (from the Greek *lithos*, meaning "stone"), which encases a hot, weak **asthenosphere** (from the Greek *asthenes*, meaning "weak"). The lithosphere includes the crust and the top part of the mantle down to an average depth of about 100 km. The asthenosphere is the portion of mantle, perhaps 300 km thick, immediately below the lithosphere. When subjected to force, the lithosphere tends to behave like a nearly rigid and brittle shell, whereas the underlying asthenosphere flows like a moldable, or *ductile*, solid.

According to the remarkable theory of *plate tectonics*, the lithosphere is not a continuous shell; it is broken into about a dozen large plates that move over Earth's surface at rates of a few centimeters per year. Each lithospheric plate is a rigid unit that rides on the asthenosphere, which is also in motion. The lithosphere that forms a plate may vary from just a few kilometers thick in volcanically active areas to more than 200 km thick beneath the older, colder parts of continents. The discovery of plate tectonics in the 1960s led to the first unified theory that explained the worldwide distribution of earthquakes and volcanoes, continental drift, mountain building, and many other geologic phenomena. Chapter 2 describes the basic concepts of plate tectonics.

Why do the plates move across Earth's surface instead of locking up into a completely rigid shell? The forces that push and pull the plates come from the mantle. Driven by Earth's internal heat engine, hot mantle material rises at boundaries where plates separate, forming new lithosphere. The lithosphere cools and becomes more rigid as it moves away from these boundaries, eventually sinking back into the mantle under the pull of gravity at other boundaries where plates converge. This general process, in which hotter material rises and cooler material sinks, is called **convection** (Figure 1.16). Convection in the mantle can

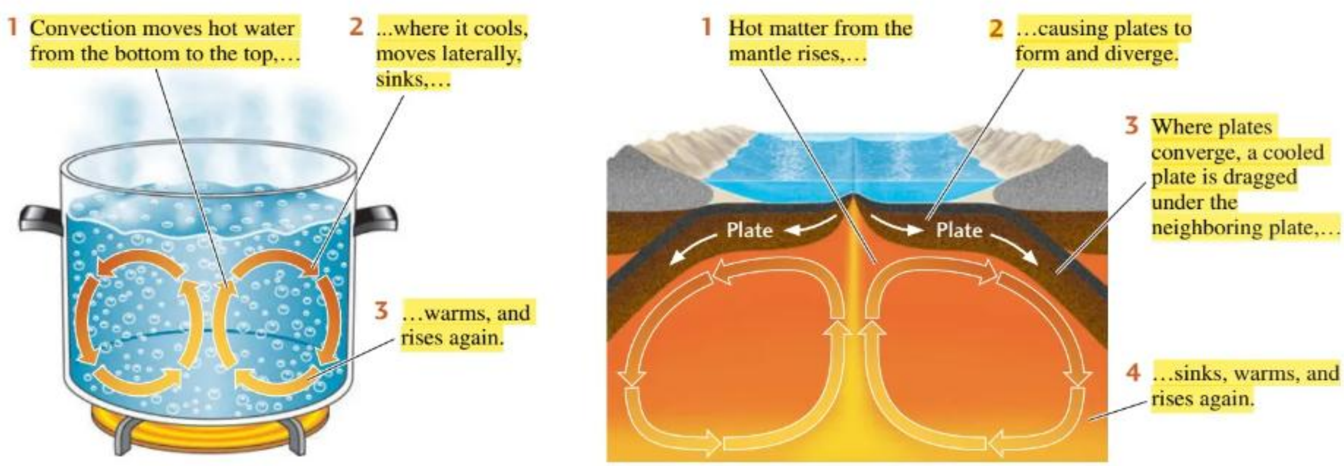


FIGURE 1.16 ■ Convection in Earth's mantle can be compared to the pattern of movement in a pot of boiling water. Both processes carry heat upward through the movement of matter.

FIGURE 2.5 ▪ Marie Tharp and Bruce Heezen inspecting a map of the seafloor. Their discovery of tectonically active rifts on mid-ocean ridges provided important evidence for seafloor spreading. [Marie Tharp, www.marietharp.com.]

The basic elements of the new theory of **plate tectonics** were established by the end of 1968. By 1970, the evidence for plate tectonics had become so persuasive that almost all Earth scientists embraced the theory. Textbooks were revised, and specialists began to consider the implications of the new concept for their own fields.

The Plates and Their Boundaries

According to the theory of plate tectonics, the lithosphere is not a continuous shell, but is broken into a mosaic of rigid plates that move over Earth's surface (Figure 2.7). Each plate travels as a distinct unit, riding on the asthenosphere, which is also in motion. The largest is the Pacific Plate, which comprises much (though not all) of the Pacific Ocean basin. Some of the plates are named after the continents they include, but in no case is a plate identical with a continent. The North American Plate, for instance, extends from the Pacific coast of North America to the middle of the Atlantic Ocean, where it meets the Eurasian and African plates.

In addition to the 13 major plates, there are a number of smaller ones. An example is the Juan de Fuca Plate, a small piece of oceanic lithosphere trapped between the giant Pacific and North American plates just offshore of the northwestern United States. Others are continental fragments, such as the small Anatolian Plate, which includes much of Turkey.

To see plate tectonics in action, go to a plate boundary. Depending on which boundary you visit, you may find earthquakes, volcanoes, rising mountains, long, narrow rifts, folding, or faulting. Many geologic features develop through the interactions of plates at their boundaries.

There are three basic types of plate boundaries (Figure 2.8), all defined by the direction of movement of the plates relative to each other:

- At **divergent boundaries,** plates move apart and new lithosphere is created (plate area increases).
- At **convergent boundaries,** plates come together and one plate is recycled into the mantle (plate area decreases).
- At **transform faults,** plates slide horizontally past each other (plate area does not change).

Like many models of nature, these three plate boundary types are idealized. There are also "oblique" boundaries that combine divergence or convergence with some amount of transform faulting. Moreover, what actually goes on at a plate boundary depends on the type of lithosphere involved, because continental and oceanic lithosphere

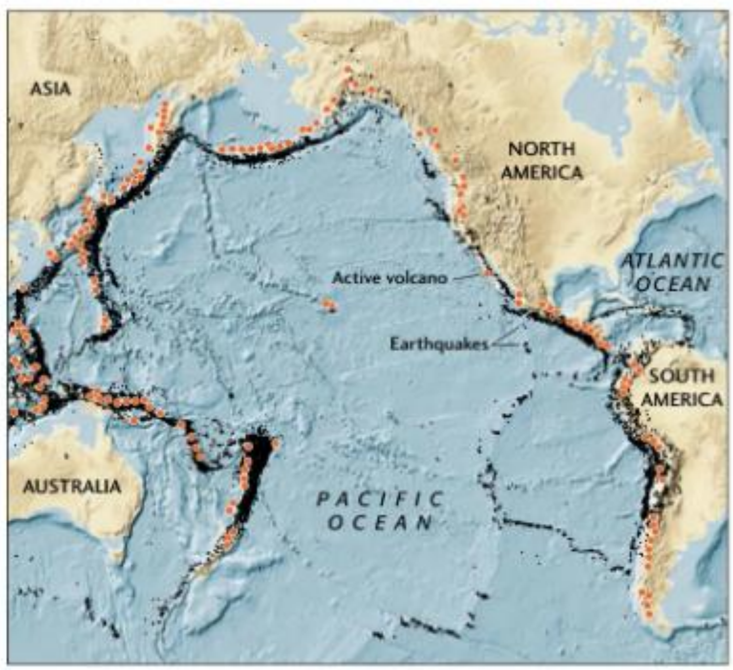


FIGURE 2.6 ▪ The Pacific Ring of Fire, with its active volcanoes (large red circles) and frequent earthquakes (small black dots), marks convergent plate boundaries where oceanic lithosphere is being recycled.

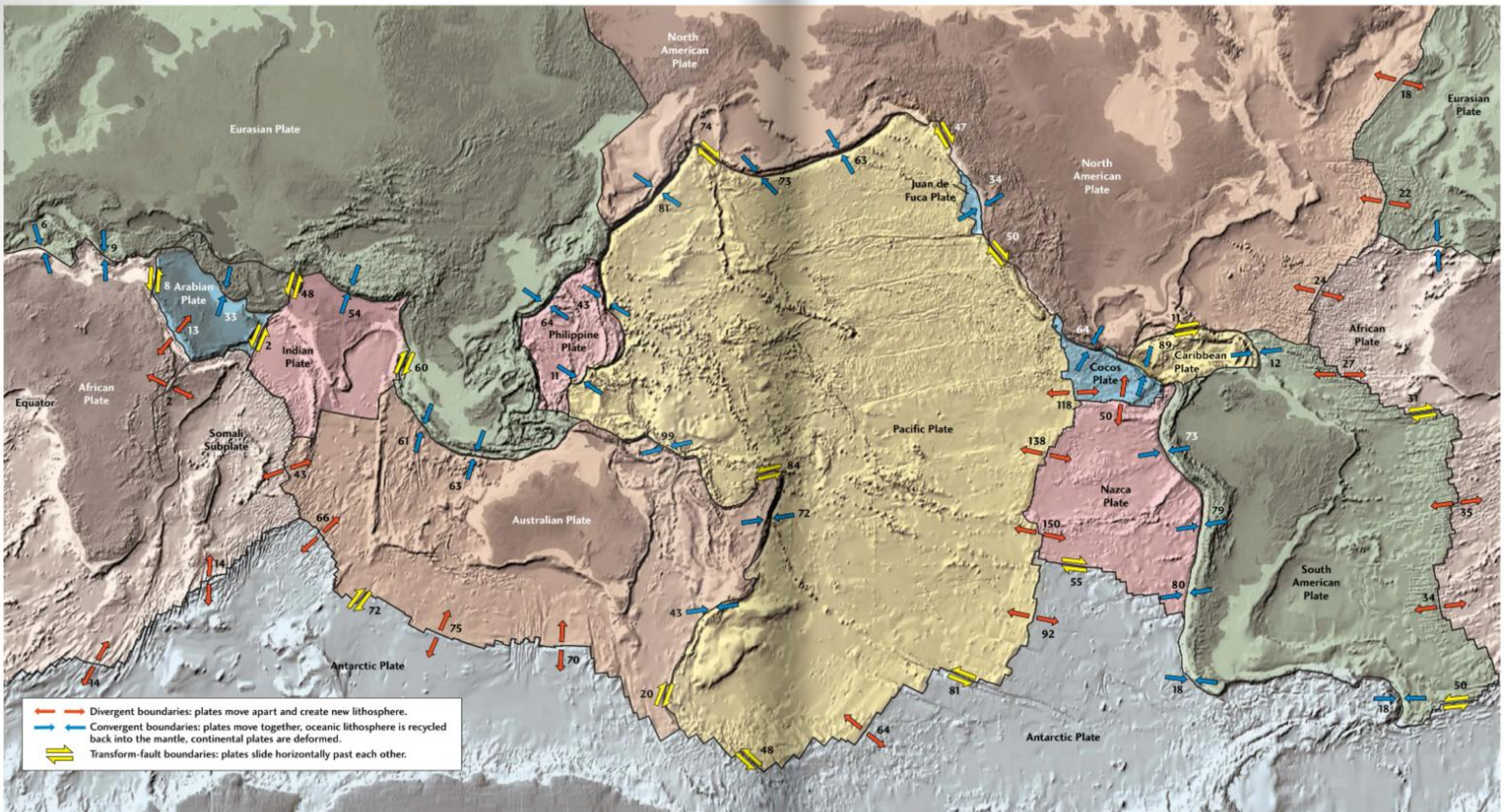


FIGURE 2.7 ▪ Earth's surface is a mosaic of 13 major plates, as well as a number of smaller plates, of rigid lithosphere that move slowly over the ductile asthenosphere. Only one of the smaller plates—the Juan de Fuca Plate, off the west coast of North America—is shown on this map. The arrows show the relative movement of two plates at a point on their boundary. The numbers next to the arrows give the relative plate velocities in millimeters per year. [Plate boundaries by Peter Bird, UCLA.]

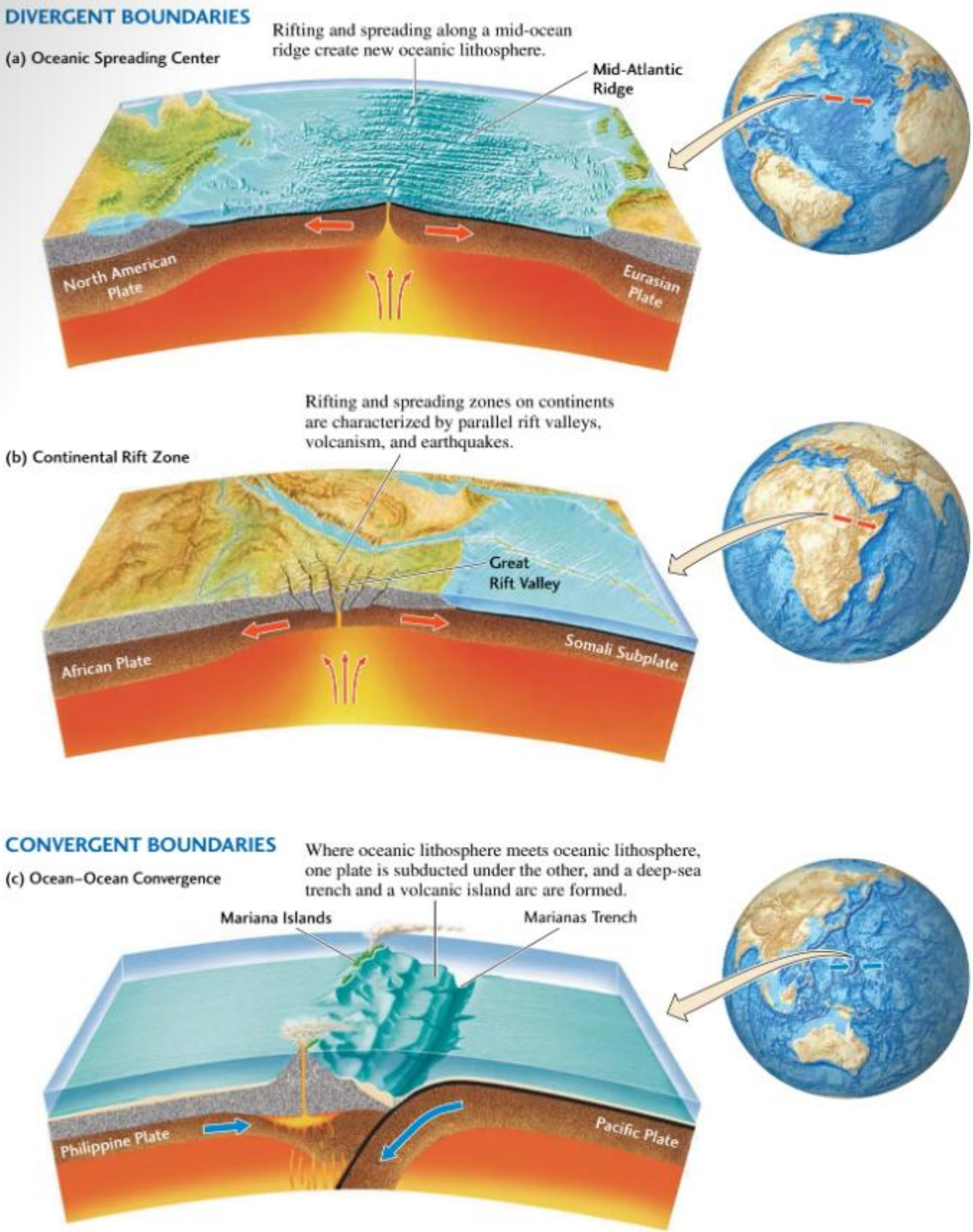
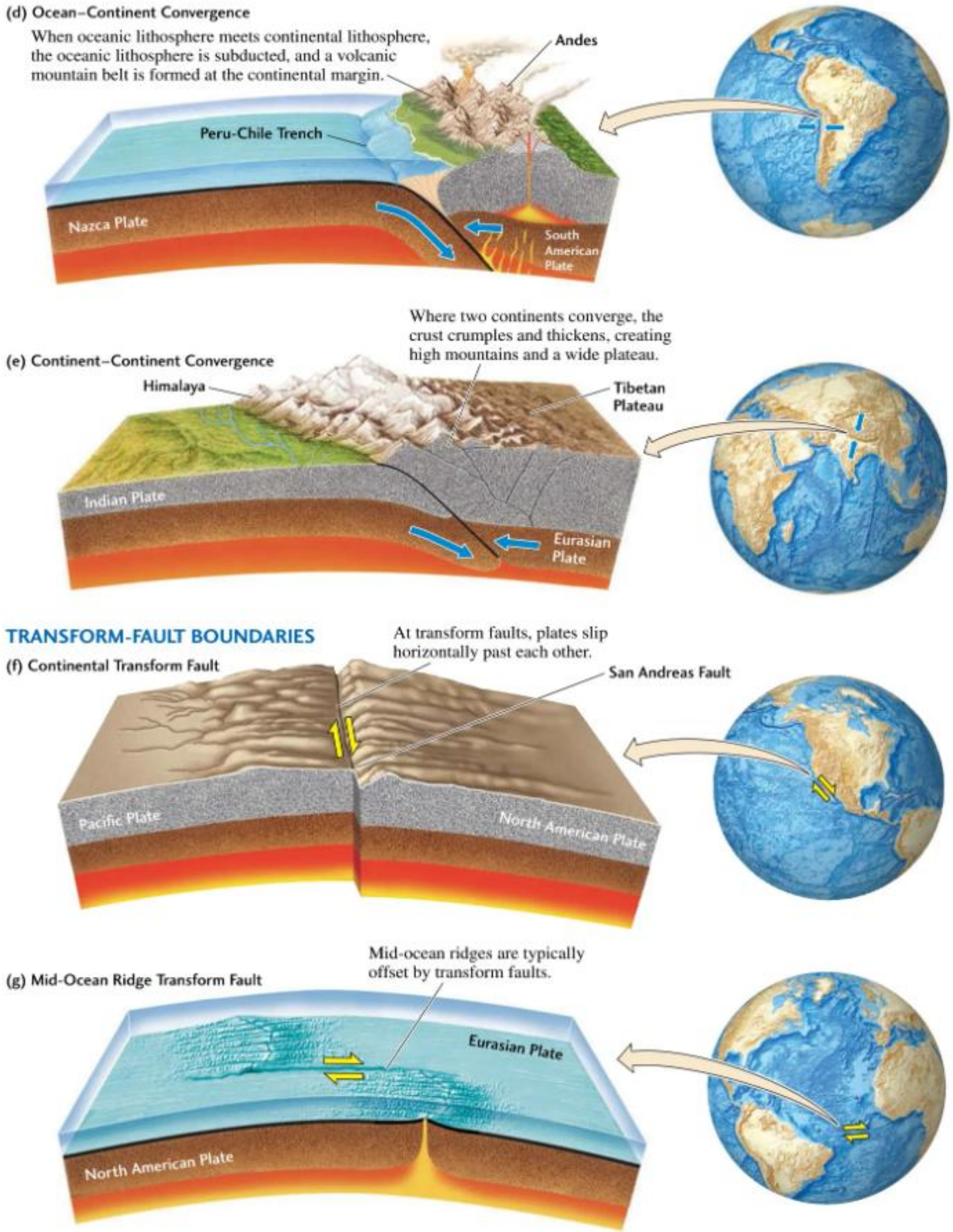


FIGURE 2.8 ▪ The interactions of lithospheric plates at their boundaries depend on the relative direction of plate movement and the type of lithosphere involved.


(d) Ocean–Continent Convergence
When oceanic lithosphere meets continental lithosphere, the oceanic lithosphere is subducted, and a volcanic mountain belt is formed at the continental margin.
Andes
Peru-Chile Trench
Nazca Plate
South American Plate
Where two continents converge, the crust crumples and thickens, creating high mountains and a wide plateau.
(e) Continent–Continent Convergence
Himalaya
Tibetan Plateau
Indian Plate
Eurasian Plate
TRANSFORM-FAULT BOUNDARIES
At transform faults, plates slip horizontally past each other.
(f) Continental Transform Fault
San Andreas Fault
Pacific Plate
North American Plate
Mid-ocean ridges are typically offset by transform faults.
(g) Mid-Ocean Ridge Transform Fault
Eurasian Plate
North American Plate