Bedrock exposed at Hallett Cove, along the south coast of Australia, an area that lies in a warm climate now, was scratched by glaciers 300 million years ago.

Chapter Objectives

By the end of this chapter you should know . . .

- Wegener's evidence for continental drift.
- how study of paleomagnetism proves that continents move.
- how sea-floor spreading works, and how geologists can prove that it takes place.
- that the Earth's lithosphere is divided into about 20 plates that move relative to one another.
- the three kinds of plate boundaries and the basis for recognizing them.
- how fast plates move, and how we can measure the rate of movement.

We are like a judge confronted by a defendant who declines to answer, and we must determine the truth from the circumstantial evidence.

—Alfred Wegener (German scientist, 1880–1930; on the challenge of studying the Earth)

2.1 Introduction

In September 1930, fifteen explorers led by a German meteorologist, Alfred Wegener, set out across the endless snowfields of Greenland to resupply two weather observers stranded at a remote camp. The observers had been planning to spend the long polar night recording wind speeds and temperatures on Greenland's polar plateau. At the time, Wegener was well known, not only to researchers studying climate but also to geologists. Some fifteen years earlier, he had published a small book, *The Origin of the Continents and Oceans*, in which he had dared to challenge geologists' long-held assumption that the continents had remained fixed in position through all of Earth history. Wegener thought, instead, that the continents once fit together like pieces of a giant jigsaw puzzle, to make one vast supercontinent. He suggested that this supercontinent, which he named **Pangaea** (pronounced pan-Jee-ah; Greek for all land), later fragmented into separate continents that drifted apart, moving slowly to their present positions (**Fig. 2.1a, b**). This model came to be known as **continental drift**.

Wegener presented many observations in favor of continental drift, but he met with strong resistance. At a widely publicized 1926 geology conference in New York City, a crowd of celebrated American professors scoffed, "What force could possibly be great enough to move the immense mass of a continent?" Wegener's writings didn't provide a good answer, so despite all the supporting observations he had provided, most of the meeting's participants rejected continental drift.

Now, four years later, Wegener faced his greatest challenge. On October 30, 1930, Wegener reached the observers and dropped off enough supplies to last the winter. Wegener and one companion set out on the return trip the next day, but they never made it home.

Had Wegener survived to old age, he would have seen his hypothesis become the foundation of a scientific revolution. Today, geologists accept many aspects of Wegener's ideas and take for granted that the map of the Earth constantly changes; continents waltz around this planet's surface, variously combining and breaking apart through geologic time. The revolution began in 1960, when an American geologist, Harry Hess, proposed that as continents drift apart, new ocean floor forms between them by a process that his contemporary, Robert Dietz, also had described and named **sea-floor spreading**. Hess and others suggested that continents move toward each other when the old ocean floor between them sinks back down into the Earth's interior, a process now called **subduction**. By 1968, geologists had developed a fairly complete model encompassing continental drift, sea-floor spreading, and subduction. In this model, Earth's lithosphere, its outer, relatively rigid shell, consists of about twenty distinct pieces, or plates, that slowly move relative to each other. Because we can confirm this model using many observations, it has gained the status of a theory, which we now call the theory of plate tectonics, or simply **plate tectonics**, from the Greek word *tekton*, which means builder; plate movements "build" regional geologic features. Geologists view plate tectonics as the grand unifying theory of geology, because it can successfully explain a great many geologic phenomena.

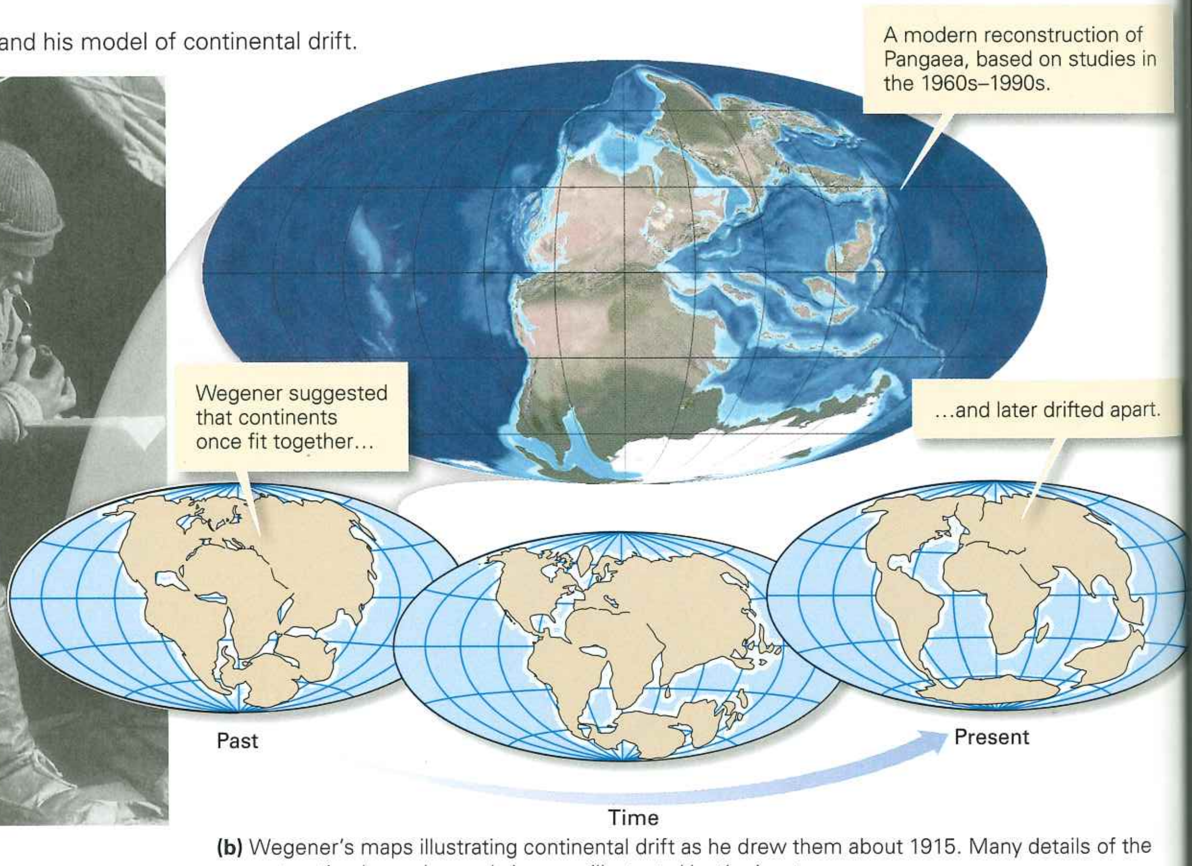
In this chapter, we introduce the observations that led Wegener to propose continental drift. Then we look at paleomagnetism, the record of Earth's magnetic field in the past, because it provides a key proof of continental drift. Next, we learn how observations about the sea floor, made by geologists during the mid-twentieth century, led Harry Hess to propose the concept of sea-floor spreading. We conclude by describing the many facets of modern plate tectonics theory.

2.2 Wegener's Evidence for Continental Drift

Wegener suggested that a vast supercontinent, Pangaea, existed until near the end of the Mesozoic Era (the interval of geologic time that lasted from 251 to 65 million years ago). He

FIGURE 2.1 Alfred Wegener and his model of continental drift.

(a) Wegener in Greenland



(b) Wegener's maps illustrating continental drift as he drew them about 1915. Many details of the reconstruction have changed since, as illustrated by the inset.

suggested that Pangaea then broke apart, and the landmasses moved away from each other to form the continents we see today. Let's look at some of Wegener's arguments and see what led him to formulate this hypothesis of continental drift.

The Fit of the Continents

Almost as soon as maps of the Atlantic coastlines became available in the 1500s, scholars noticed the fit of the continents. The northwestern coast of Africa could tuck in against the eastern coast of North America, and the bulge of eastern South America could nestle cozily into the indentation of southwestern Africa. Australia, Antarctica, and India could all connect to the southeast of Africa, while Greenland, Europe, and Asia could pack against the northeastern margin of North America. In fact, all the continents could be joined, with remarkably few overlaps or gaps, to create Pangaea. Wegener concluded that the fit was too good to be coincidence and thus that the continents once *did* fit together.

Did you ever wonder . . .
why opposite coasts of the Atlantic look like they fit together?

Locations of Past Glaciations

Glaciers are rivers or sheets of ice that flow across the land surface. As a glacier flows, it carries sediment grains of all sizes (clay, silt, sand, pebbles, and boulders). Grains protruding from the base of the moving ice carve scratches, called striations, into the substrate. When the ice melts, it leaves the sediment in a deposit called till, that buries striations. Thus, the occurrence of till and striations at a location serve as evidence that the region was covered by a glacier in the past (see chapter opening photo). By studying the age of glacial till deposits, geologists have determined that large areas of the land were covered by glaciers during time intervals of Earth history called ice ages. One of these ice ages occurred from about 326 to 267 Ma, near the end of the Paleozoic Era.

Wegener was an Arctic climate scientist by training, so it's no surprise that he had a strong interest in glaciers. He knew that glaciers form mostly at high latitudes today. So he suspected that if he plotted a map of the locations of late Paleozoic glacial till and striations, he might gain insight into the locations of continents during the Paleozoic. When he plotted these locations, he found that glaciers of this time interval occurred in southern South America, southern Africa, southern India, Antarctica, and southern Australia. These places are now widely separated and, with the exception of Antarctica, do not currently lie in cold polar regions (**Fig. 2.2a**). To Wegener's amazement, all late Paleozoic glaciated areas lie *adjacent* to each other on his map of Pangaea. Furthermore, when he plotted the orientation of glacial striations, they all pointed roughly outward from a location in southeastern Africa. In other words, Wegener determined that the distribution of glaciations at the end of the Paleozoic Era could easily be explained if the continents had been united in Pangaea, with the southern part of Pangaea lying beneath the center of a huge ice cap. This distribution of glaciation could not be explained if the continents had always been in their present positions.

The Distribution of Climatic Belts

If the southern part of Pangaea had straddled the South Pole at the end of the Paleozoic Era, then during this same time interval, southern North America, southern Europe, and northwestern Africa would have straddled the equator and would have had tropical or subtropical climates. Wegener searched for evidence that this was so by studying sedimentary rocks that were formed at this time, for the material making up these rocks can reveal clues to the past climate. For example, in the swamps and jungles of tropical regions, thick deposits of plant material accumulate, and when deeply buried, this material transforms into coal. And in the clear, shallow seas of tropical regions, large reefs develop. Finally, subtropical regions, on either side of the tropical belt, contain deserts, an environment in which sand dunes form and salt from evaporating seawater or salt lakes accumulates. Wegener speculated that the distribution of late Paleozoic coal, reef, sand-dune, and salt deposits could define climate belts on Pangaea.

Sure enough, in the belt of Pangaea that Wegener expected to be equatorial, late Paleozoic sedimentary rock layers include abundant coal and the relicts of reefs. And in the portions of Pangaea that Wegener predicted would be subtropical, late Paleozoic sedimentary rock layers include relicts of desert dunes and deposits of salt (**Fig. 2.2b**). On a present-day map of our planet, exposures of these ancient rock layers scatter around the globe at a variety of latitudes. On Wegener's Pangaea, the exposures align in continuous bands that occupy appropriate latitudes.

The Distribution of Fossils

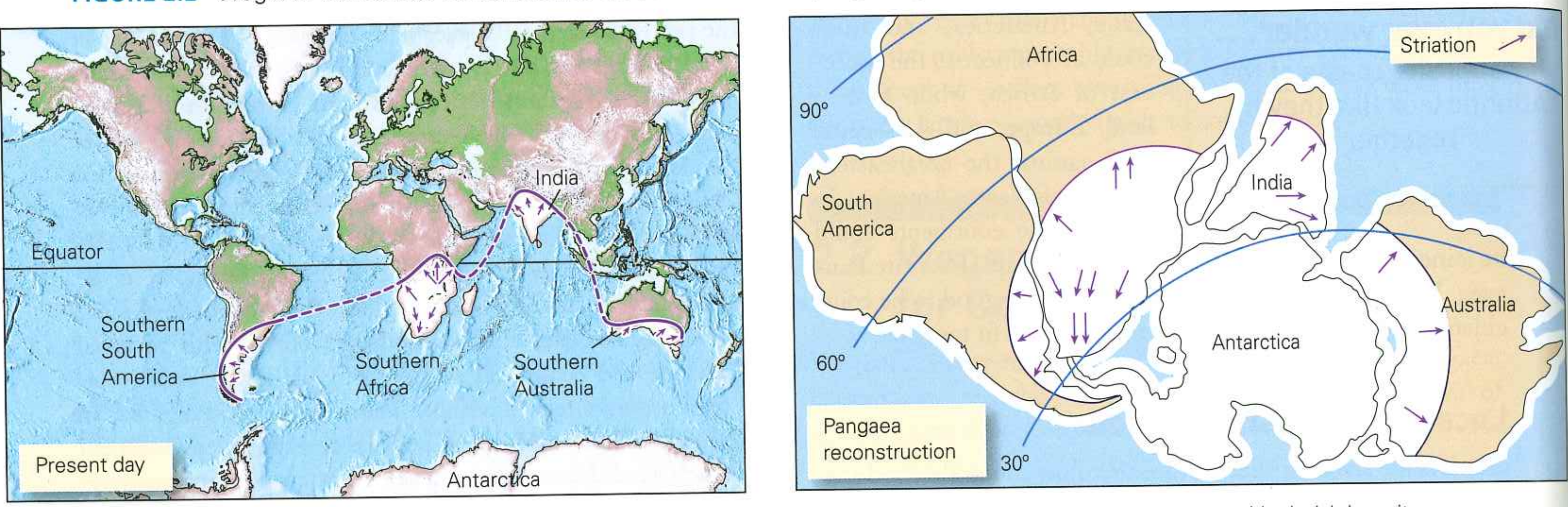
Today, different continents provide homes for different species. Kangaroos, for example, live only in Australia. Similarly, many kinds of plants grow only on one continent and not on others. Why? Because land-dwelling species of animals and plants cannot swim across vast oceans, and thus evolved independently on different continents. During a period of Earth history when all continents were in contact, however, land animals and plants could have migrated among many continents.

With this concept in mind, Wegener plotted fossil occurrences of land-dwelling species that existed during the late Paleozoic and early Mesozoic Eras (between about 300 and 210 million years ago) and found that these species had indeed existed on several continents (**Fig. 2.2c**). Wegener argued that the distribution of fossil species required the continents to have been adjacent to one another in the late Paleozoic and early Mesozoic Eras.

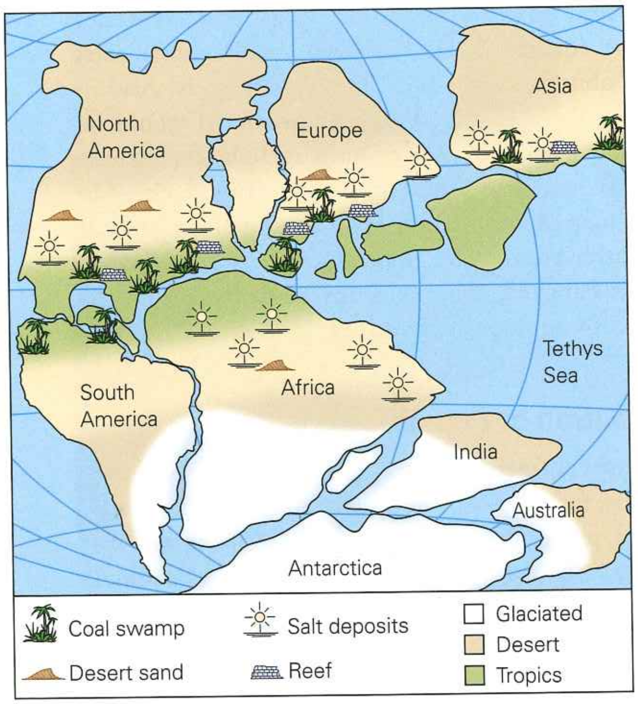
Matching Geologic Units

An art historian can recognize a Picasso painting, an architect knows what makes a building look "Victorian," and a geoscientist can identify a distinctive assemblage of rocks. Wegener found that the same distinctive Precambrian rock assemblages occurred on the eastern coast of South America and the western coast of Africa, regions now separated by an ocean (**Fig. 2.3a**). If the continents had been joined to create Pangaea in the past, then these matching rock groups would have been adjacent to each other, and thus could have composed

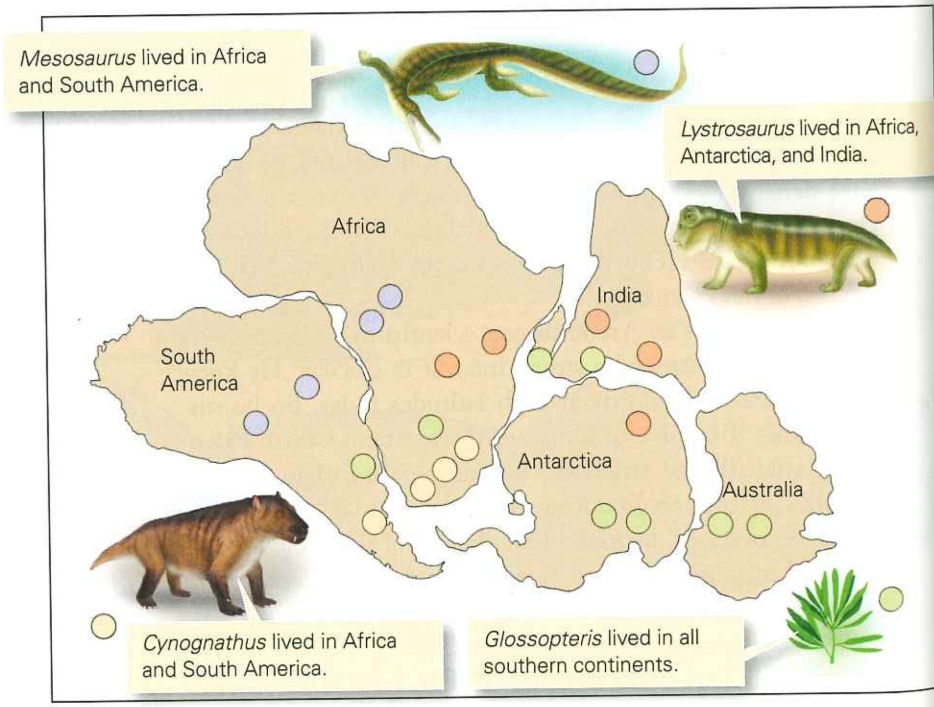
FIGURE 2.2 Wegener's evidence for continental drift came from analyzing the geologic record.



(a) The distribution of late Paleozoic glacial deposits and striations on present-day Earth are hard to explain. But on Pangaea, areas with glacial deposits fit together in a southern polar cap.



(b) The distribution of late Paleozoic rock types plots sensibly in the climate belts of Pangaea.



(c) A plot of fossil localities shows that Mesozoic land-dwelling organisms occur on multiple continents. This would be hard to explain if continents were separated.

continuous blocks or belts. Wegener also noted that features of the Appalachian Mountains of the United States and Canada closely resemble mountain belts in southern Greenland, Great Britain, Scandinavia, and northwestern Africa (**Fig. 2.3b, c**), regions that would have lain adjacent to each other in Pangaea. Wegener thus demonstrated that not only did the coastlines of continents match, so too did the rocks adjacent to the coastlines.

Criticism of Wegener's Ideas

Wegener's model of a supercontinent that later broke apart explained the distribution of ancient glaciers, coal, sand dunes, rock assemblages, and fossils. Clearly, he had compiled a strong *circumstantial* case for continental drift. But as noted earlier, he could not adequately explain how or why continents drifted. He left on his final expedition to Greenland having failed to convince his peers, and he died without knowing that his ideas, after lying dormant for decades, would be reborn as the basis of the broader theory of plate tectonics.

In effect, Wegener was ahead of his time. It would take three more decades of research before geologists obtained sufficient data to test his hypotheses properly. Collecting this data required instruments and techniques that did not exist in Wegener's day. Of the many geologic discoveries that ultimately opened the door to plate tectonics, perhaps the most important came from the discovery of a phenomenon called paleomagnetism, so we discuss it next.

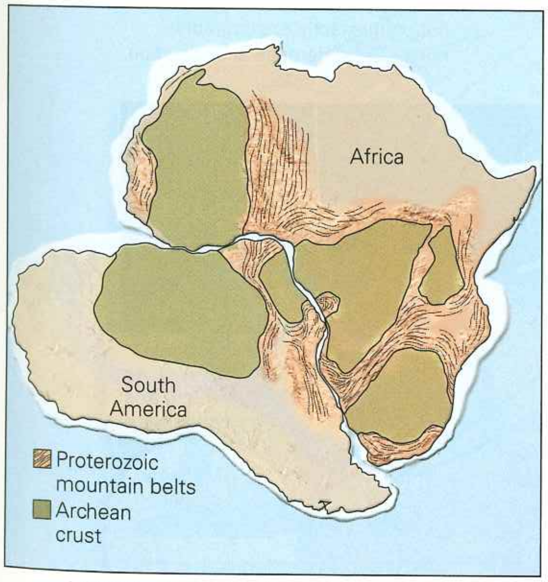
Take-Home Message

Wegener argued that the continents were once merged into a supercontinent called Pangaea that later broke up to produce smaller continents that "drifted" apart. The matching shapes of coastlines, as well as the distribution of ancient climate belts, fossils, and rock units all make better sense if Pangaea existed.

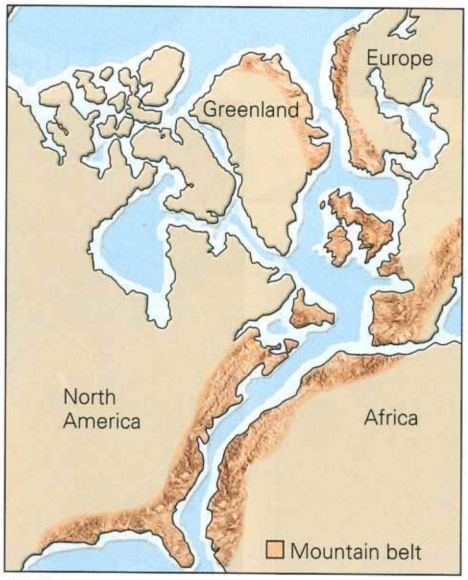
2.3 Paleomagnetism and the Proof of Continental Drift

More than 1,500 years ago, Chinese sailors discovered that a piece of lodestone, when suspended from a thread, points in a northerly direction and can help guide a voyage. Lodestone exhibits this behavior because it consists of magnetite, an iron-rich mineral that, like a compass needle, aligns with Earth's magnetic field lines. While not as magnetic as lodestone, several other rock types contain tiny crystals of magnetite, or other magnetic minerals, and thus behave overall like weak magnets. In this section, we explain how the study of such magnetic behavior led to the realization that rocks preserve paleomagnetism, a record of Earth's magnetic field in the past. An understanding of paleomagnetism provided proof of continental drift and, as we'll see later in this chapter, contributed

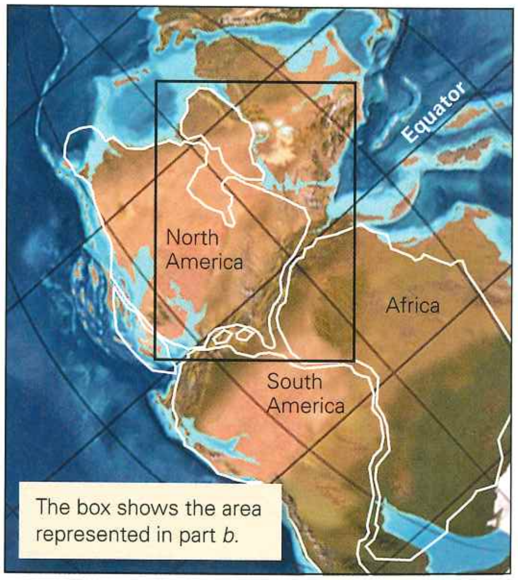
FIGURE 2.3 Further evidence of drift: rocks on different sides of the ocean match.



(a) Distinctive belts of rock in South America would align with similar ones in Africa, without the Atlantic.



(b) If the Atlantic didn't exist, Paleozoic mountain belts on both coasts would be adjacent.



(c) A modern reconstruction showing the positions of mountain belts in Pangaea. Modern continents are outlined in white.

to the development of plate tectonics theory. As a foundation for introducing paleomagnetism, we first provide additional detail about the basic nature of the Earth's magnetic field.

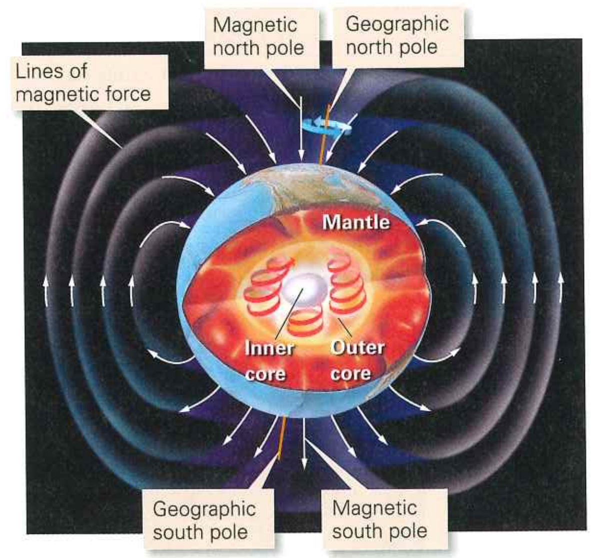
Earth's Magnetic Field

As we mentioned in Chapter 1, circulation of liquid iron alloy in the outer core of the Earth generates a magnetic field. (A similar phenomenon happens in an electrical dynamo at a power plant.) Earth's magnetic field resembles the field produced by a bar magnet, in that it has two ends of opposite polarity. Thus, we can represent Earth's field by a **magnetic dipole**, an imaginary arrow (Fig. 2.4a). Earth's dipole intersects the surface of the planet at two points, known as the **magnetic poles**. By convention, the north *magnetic* pole is at the end of the Earth nearest the north *geographic* pole (the point where the northern end of the spin axis intersects the surface). The north-seeking (red) end of a compass needle points to the north magnetic pole.

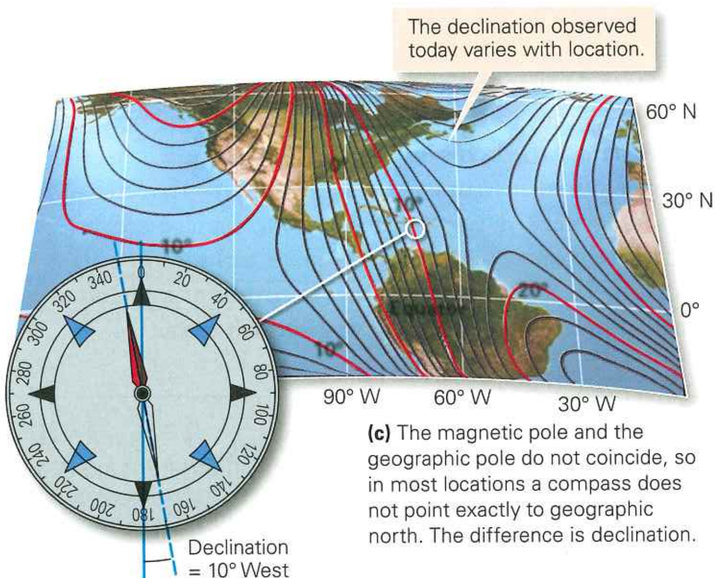
Did you ever wonder . . .
why compasses always point to the north?

Earth's magnetic poles move constantly, but don't seem to stray further than about 1,500 km from the geographic poles, and averaged over thousands of years, they roughly coincide with Earth's geographic poles (Fig. 2.4b). That's because the rotation of the Earth causes the flow to organize into patterns resembling spring-like spirals, and these are roughly aligned with the spin axis. At present, the magnetic poles lie hundreds of kilometers away from the geographic poles, so the magnetic dipole tilts at about 11° relative to the Earth's spin axis. Because of this difference, a compass today does not point exactly to geographic north. The angle between the direction that a compass needle points and a line of longitude at a given location is the **magnetic declination** (Fig. 2.4c).

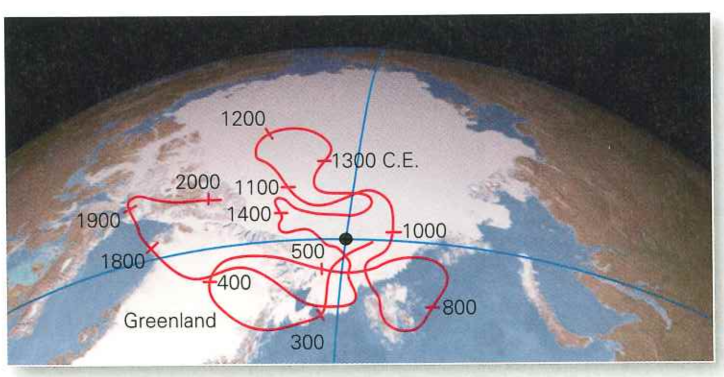
FIGURE 2.4 Features of Earth's magnetic field.



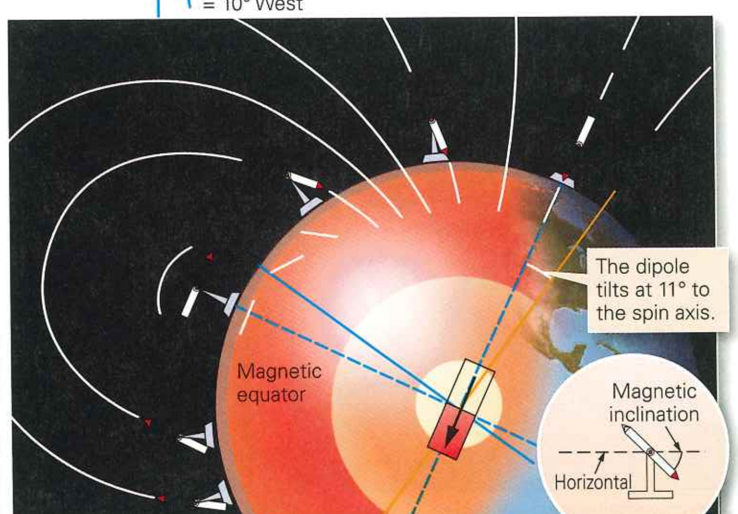
(a) The magnetic axis is not parallel to the spin axis. In 3-D, the Earth's magnetic field can be visualized as invisible curtains of energy, generated by flow to the outer core.



(c) The magnetic pole and the geographic pole do not coincide, so in most locations a compass does not point exactly to geographic north. The difference is declination.



(b) A map of the magnetic pole position during the past 1,800 years shows that the pole moves, but stays within high latitudes.



(d) Earth's field lines curve, so the tilt of a magnetic needle changes with latitude. This tilt is the magnetic inclination.

Invisible field lines curve through space between the magnetic poles. In a cross-sectional view, these lines lie parallel to the surface of the Earth (that is, are horizontal) at the equator, tilt at an angle to the surface in midlatitudes, and plunge perpendicular to the surface at the magnetic poles (**Fig. 2.4d**). The angle between a magnetic field line and the surface of the Earth, at a given location, is called the **magnetic inclination**. If you place a magnetic needle on a horizontal axis so that it can pivot up and down, and then carry it from the magnetic equator to the magnetic pole, you'll see that the inclination varies with latitude—it is 0° at the magnetic equator and 90° at the magnetic poles. (Note that the compass you may carry with you on a hike does not show inclination because it has been balanced to remain horizontal.)

What Is Paleomagnetism?

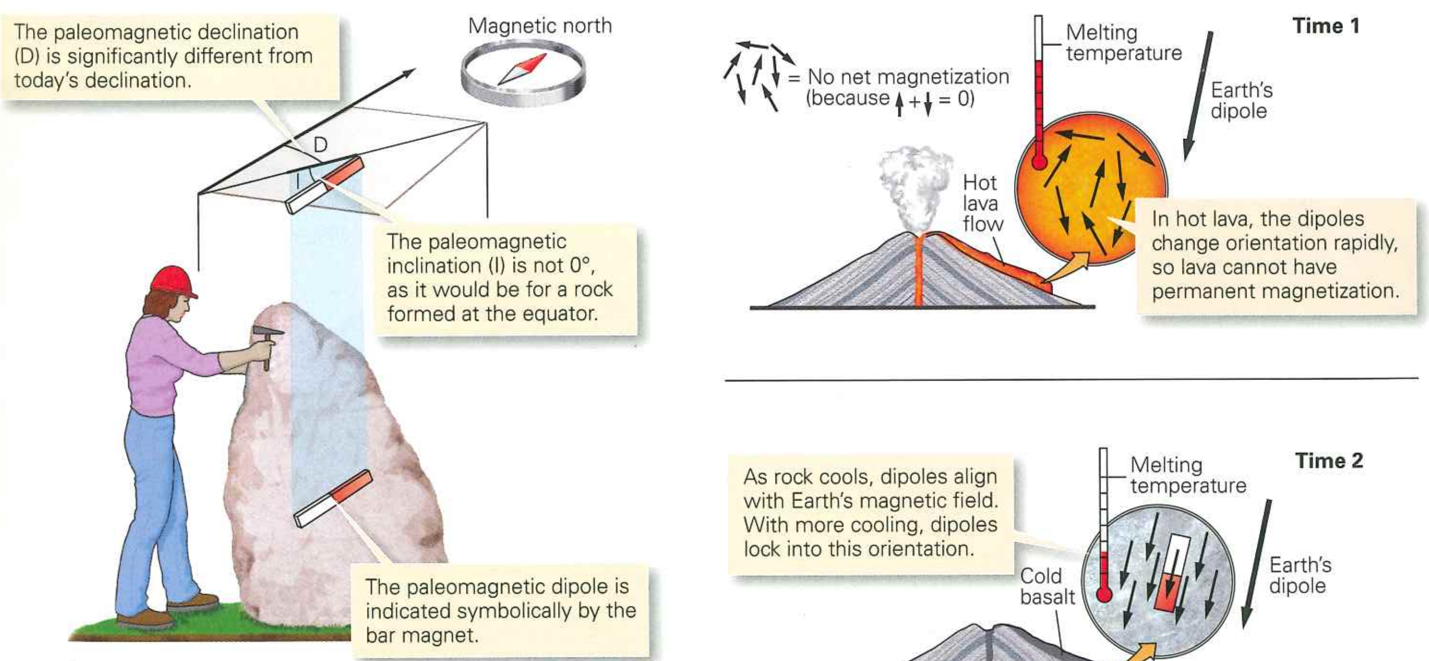
In the early 20th century, researchers developed instruments that could measure the weak magnetic field produced by rocks and made a surprising discovery. In a rock that formed millions of years ago, the orientation of the dipole representing the magnetic field of the rock *is not the same* as that of present-day Earth (**Fig. 2.5a**). To understand this statement, consider an example. Imagine traveling to a location near the coast on the equator in South America where the inclination and declination are presently 0°. If you measure the weak magnetic field produced by, say, a 90-million-year-old rock, and represent the orientation of this field by an imaginary bar magnet, you'll find that this imaginary bar magnet does *not* point to the present-day north magnetic pole, and you'll find that its inclination is not 0°. The reason for this difference is that the magnetic fields of ancient rocks indicate the orientation of the magnetic field, relative to the rock, at the time the rock formed. This record, preserved in rock, is **paleomagnetism**.

Paleomagnetism can develop in many different ways. For example, when lava, molten rock containing no crystals, starts to cool and solidify into rock, tiny magnetite crystals begin to grow (**Fig. 2.5b**). At first, thermal energy causes the tiny magnetic dipole associated with each crystal to wobble and tumble chaotically. Thus, at any given instant, the dipoles of the magnetite specks are randomly oriented and the magnetic forces they produce cancel each other out. Eventually, however, the rock cools sufficiently that the dipoles slow down and, like tiny compass needles, align with the Earth's magnetic field. As the rock cools still more, these tiny compass needles lock into permanent parallelism with the Earth's magnetic field at the time the cooling takes place. Since the magnetic dipoles of all the grains point in the same direction, they add together and produce a measurable field.

Apparent Polar Wander—A Proof That Continents Move

Why doesn't the paleomagnetic dipole in ancient rocks point to the present-day magnetic field? When geologists first

FIGURE 2.5 Paleomagnetism and how it can form during the solidification and cooling of lava.



(a) A geologist finds an ancient rock sample at a location on the equator, where declination today is 0°. The orientation of the rock's paleomagnetism is different from that of today's field.

(b) Paleomagnetism can form when lava cools and becomes solid rock.

attempted to answer this question, they *assumed* that continents were fixed in position and thus concluded that the positions of Earth's magnetic poles in the past were different than they are today. They introduced the term **paleopole** to refer to the supposed position of the Earth's magnetic north pole in the past. With this concept in mind, they set out to track what they thought was the change in position of the paleopole over time. To do this, they measured the paleomagnetism in a succession of rocks of different ages from the same general location on a continent, and they plotted the position of the associated succession of paleopole positions on a map (**Fig. 2.6a**). The successive positions of dated paleopoles trace out a curving line that came to be known as an **apparent polar-wander path.**

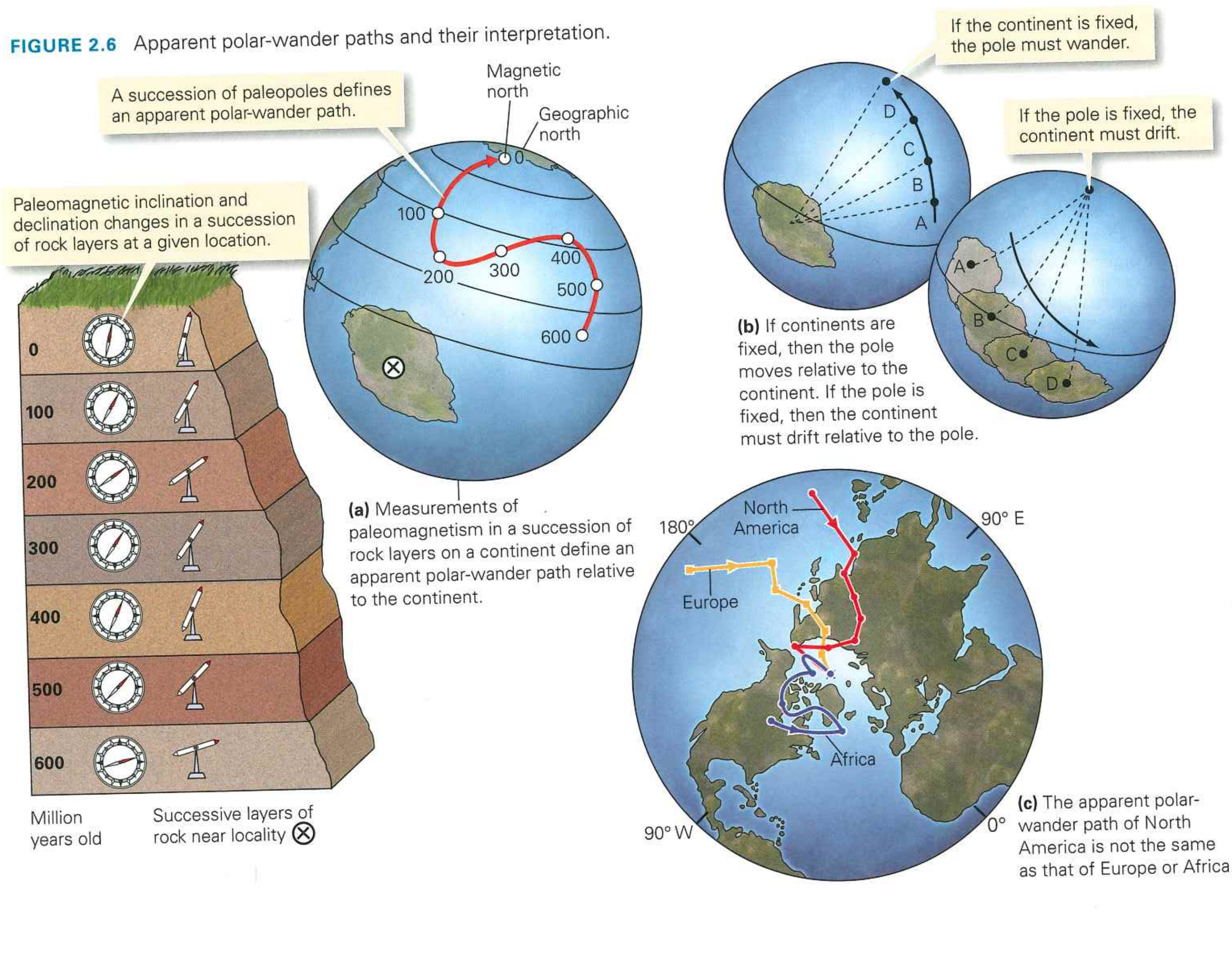
At first, geologists assumed that the apparent polar-wander path actually represented how the position of Earth's magnetic pole migrated through time. But were they in for a surprise! When they obtained polar-wander paths from many different continents, they found that *each continent has a different apparent polar-wander path*. The hypothesis that continents are fixed in position cannot explain this observation, for if the magnetic pole moved while all the continents stayed fixed, measurements from all continents should produce the same apparent polar-wander paths.

Geologists suddenly realized that they were looking at apparent polar-wander paths in the wrong way. It's not the pole that moves relative to fixed continents, but rather the continents that move relative to a fixed pole (**Fig. 2.6b**). Since each continent has its own unique polar-wander path (**Fig. 2.6c**), *the continents must move with respect to each other.* The discovery proved that Wegener was essentially right all along—continents do move!

Take-Home Message

Study of paleomagnetism indicates that the continents have moved relative to the Earth's magnetic poles. Each continent has a different apparent polar-wander path, which is only possible if the continents move ("drift") relative to each other.

FIGURE 2.6 Apparent polar-wander paths and their interpretation.



(a) Measurements of paleomagnetism in a succession of rock layers on a continent define an apparent polar-wander path relative to the continent.

(b) If continents are fixed, then the pole moves relative to the continent. If the pole is fixed, then the continent must drift relative to the pole.

(c) The apparent polar-wander path of North America is not the same as that of Europe or Africa.

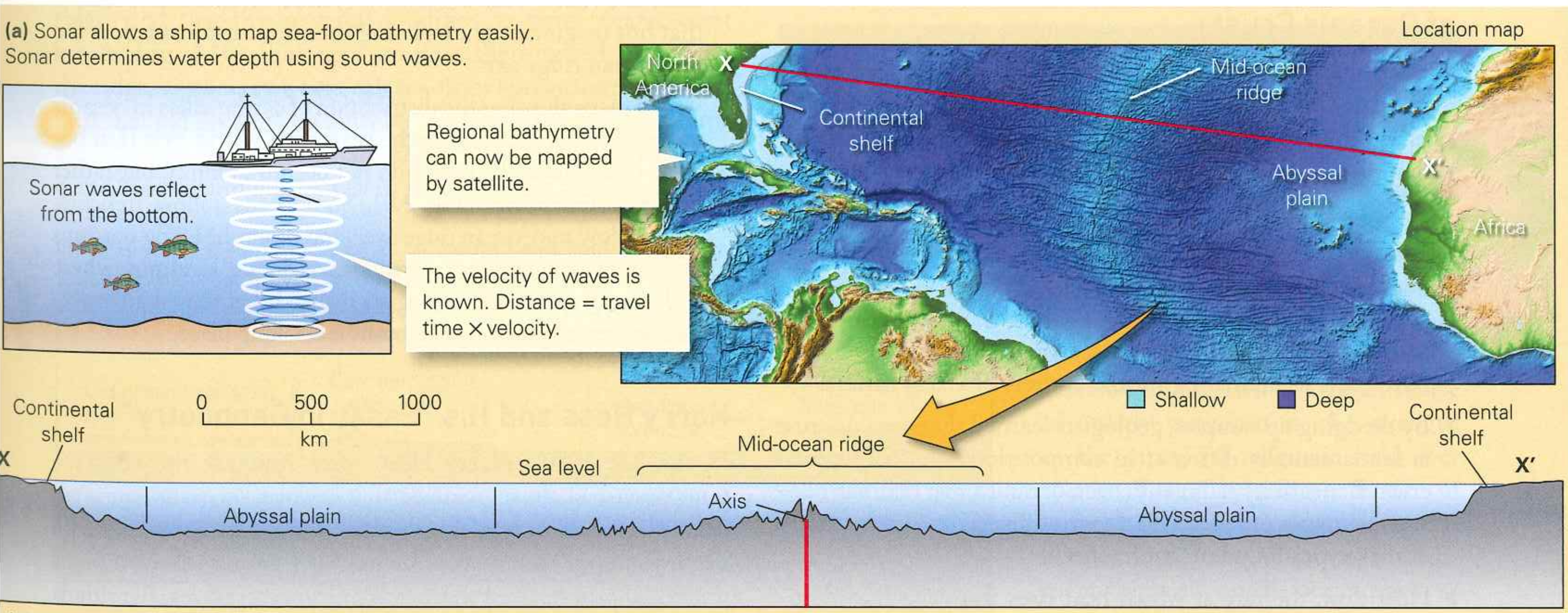
2.4 The Discovery of Sea-Floor Spreading

New Images of Sea-Floor Bathymetry

Military needs during World War II gave a boost to sea-floor exploration, for as submarine fleets grew, navies required detailed information about **bathymetry**, or depth variations. The invention of echo sounding (sonar) permitted such information to be gathered quickly. Echo sounding works on the same principle that a bat uses to navigate and find insects. A sound pulse emitted from a ship travels down through the water, bounces off the sea floor, and returns up as an echo through the water to a receiver on the ship. Since sound waves travel at a known velocity, the time between the sound emission and the echo detection indicates the distance between the ship and the sea floor. (Recall that velocity = distance/time, so distance = velocity × time.) As the ship travels, observers can obtain a continuous record of the depth of the sea floor. The resulting cross section showing depth plotted against location is called a bathymetric profile (**Fig. 2.7a, b**). By cruising back and forth across the ocean many times, investigators obtained a series of bathymetric profiles and from these constructed maps of the sea floor. (Geologists can now produce such maps much more rapidly using satellite data.) Bathymetric maps reveal several important features.

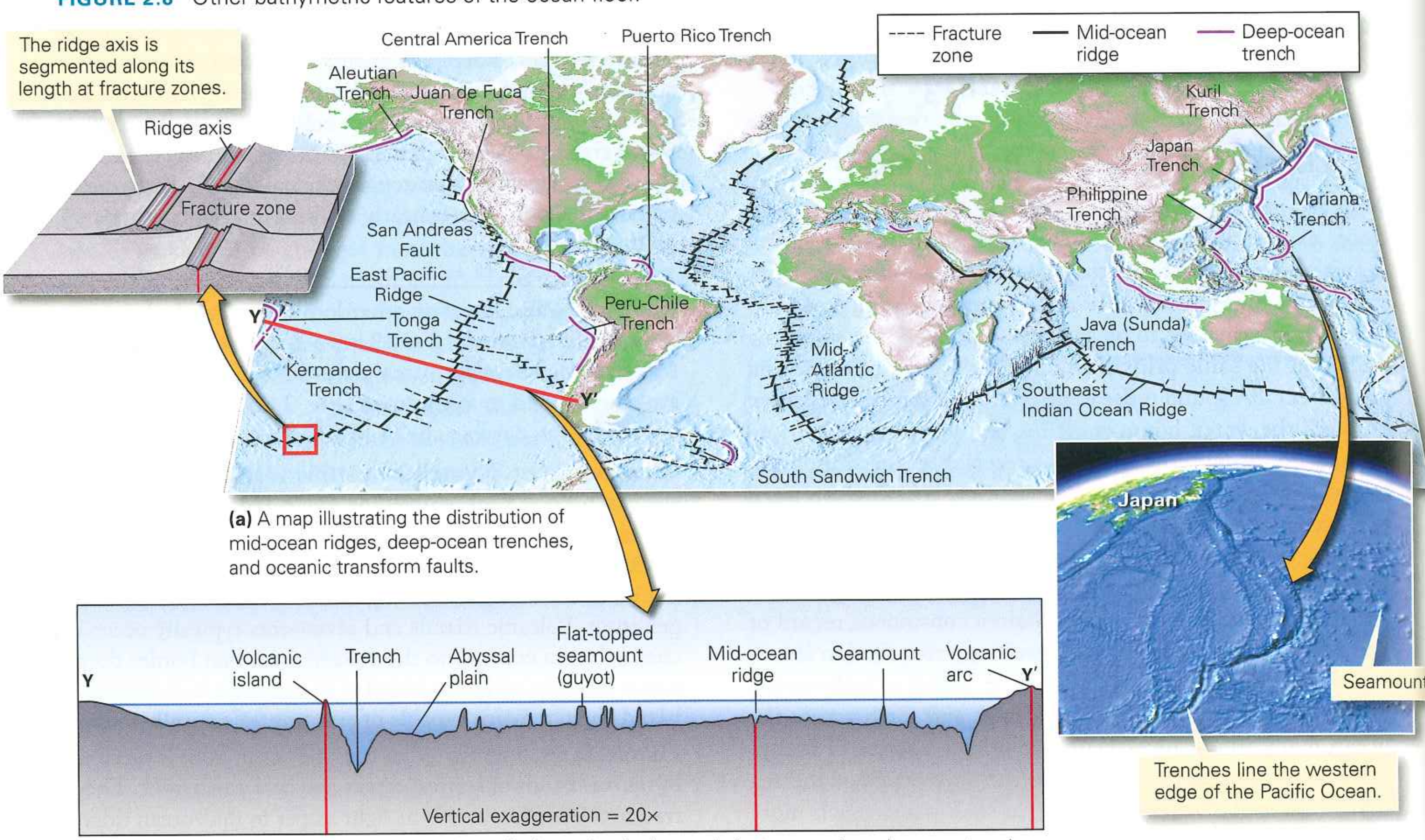
- *Mid-ocean ridges*: The floor beneath all major oceans includes **abyssal plains**, which are broad, relatively flat regions of the ocean that lie at a depth of about 4 to 5 km below sea level; and **mid-ocean ridges**, submarine mountain ranges whose peaks lie only about 2 to 2.5 km below sea level (**Fig. 2.8a**). Geologists call the crest of the mid-ocean ridge the ridge axis. All mid-ocean ridges are roughly symmetrical—bathymetry on one side of the axis is nearly a mirror image of bathymetry on the other side.
- *Deep-ocean trenches*: Along much of the perimeter of the Pacific Ocean, and in a few other localities as well, the ocean floor reaches depths of 8 to 12 km—deep enough to swallow Mt. Everest. These deep areas occur in elongate troughs that are now referred to as **trenches** (**Fig. 2.8b**). Trenches border **volcanic arcs**, curving chains of active volcanoes.
- *Seamount chains*: Numerous volcanic islands poke up from the ocean floor: for example, the Hawaiian Islands lie in the middle of the Pacific. In addition to islands that rise above sea level, sonar has detected many **seamounts** (isolated submarine mountains), which were once volcanoes but no longer erupt. Volcanic islands and seamounts typically occur in chains, but in contrast to the volcanic arcs that border deep-ocean trenches, only one island at the end of a seamount and island chain remains capable of erupting volcanically today.
- *Fracture zones*: Surveys reveal that the ocean floor is diced up by narrow bands of vertical cracks and broken-up rock. These **fracture zones** lie roughly at right angles to mid-ocean ridges. The ridge axis typically steps sideways when it intersects with a fracture zone.

FIGURE 2.7 Bathymetry of mid-ocean ridges and abyssal plains.



(b) A bathymetric profile along line X–X' illustrates how mid-ocean ridges rise above abyssal plains. Both are deeper than continental shelves.

FIGURE 2.8 Other bathymetric features of the ocean floor.



(a) A map illustrating the distribution of mid-ocean ridges, deep-ocean trenches, and oceanic transform faults.

(b) In addition to mid-ocean ridges, the sea floor displays other bathymetric features such as deep-sea trenches, oceanic islands, guyots, and seamounts, as shown in this vertically exaggerated profile.

New Observations on the Nature of Oceanic Crust

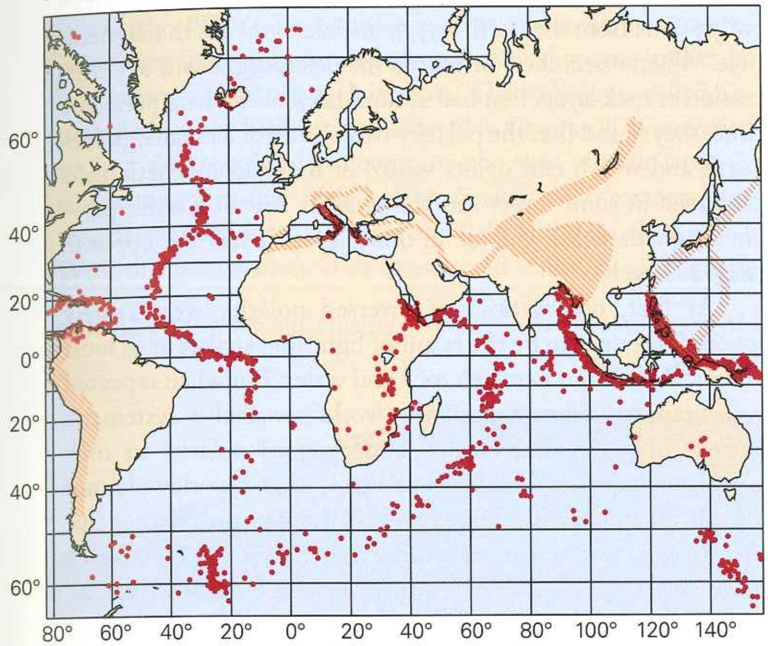
By the mid-20th century, geologists had discovered many important characteristics of the sea-floor crust. These discoveries led them to realize that oceanic crust differs from continental crust, and that bathymetric features of the ocean floor provide clues to the origin of the crust. Specifically:

- A layer of sediment composed of clay and the tiny shells of dead plankton covers much of the ocean floor. This layer becomes progressively thicker away from the mid-ocean ridge axis. But even at its thickest, the sediment layer is too thin to have been accumulating for the entirety of Earth history.
- By dredging up samples, geologists learned that oceanic crust is fundamentally different in composition from continental crust. Beneath its sediment cover, oceanic crust bedrock consists primarily of basalt—it does not display the great variety of rock types found on continents.
- Heat flow, the rate at which heat rises from the Earth's interior up through the crust, is not the same everywhere in the oceans. Rather, more heat rises beneath mid-ocean ridges than elsewhere. This observation led researchers to speculate that hot magma might be rising into the crust just below the mid-ocean ridge axis.
- When maps showing the distribution of earthquakes in oceanic regions became available in the years after World War II, it became clear that earthquakes do not occur randomly, but rather define distinct belts (**Fig. 2.9**). Some belts follow trenches, some follow mid-ocean ridge axes, and others lie along portions of fracture zones. Since earthquakes define locations where rocks break and move, geologists realized that these bathymetric features are places where motion is taking place.

Harry Hess and His "Essay in Geopoetry"

In the late 1950s, Harry Hess, after studying the observations described above, realized that because the sediment layer on the ocean floor was thin overall, the ocean floor might be much younger than the continents. Also, because the sediment thickened progressively away from mid-ocean ridges, the ridges themselves likely were younger than the deeper parts of the

FIGURE 2.9 A 1953 map showing the distribution of earthquake locations in the ocean basins. Note that earthquakes occur in belts.



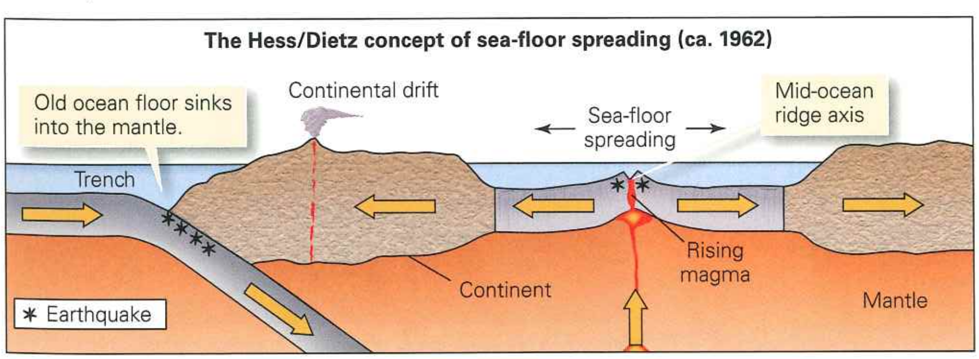
into the mantle. Hess suggested that earthquakes at trenches were evidence of this movement, but he didn't understand how the movement took place. Other geologists, such as Robert Dietz, were coming to similar conclusions at about the same time.

Did you ever wonder . . . if the distance between New York and Paris changes?

Hess and his contemporaries realized that the sea-floor-spreading hypothesis instantly provided the long-sought explanation of how continental "drift" occurs. Continents passively move apart as the sea floor between them spreads at mid-ocean ridges, and they passively move together as the sea floor between them sinks back into the mantle at trenches. (As we will see later, geologists now realize that it is the lithosphere that moves, not just the crust.) Thus, sea-floor spreading proved to be an important step on the route to plate tectonics—the idea seemed so good that Hess referred to his description of it as "an essay in geopoetry." But first, the idea needed to be tested, and other key discoveries would have to take place before the whole theory of plate tectonics could come together.

ocean floor. If this was so, then somehow *new ocean floor must be forming at the ridges*, and thus an ocean basin could be getting wider with time. But how? The association of earthquakes with mid-ocean ridges suggested to him that the sea floor was cracking and splitting apart at the ridge. The discovery of high heat flow along mid-ocean ridge axes provided the final piece of the puzzle, for it suggested the presence of very hot molten rock beneath the ridges. In 1960, Hess suggested that indeed molten rock (basaltic magma) rose upward beneath mid-ocean ridges and that this material solidified to form oceanic crust basalt (**Fig. 2.10**). The new sea floor then moved away from the ridge, a process we now call **sea-floor spreading**. Hess realized that old ocean floor must be consumed somewhere, or the Earth would have to be expanding, so he suggested that deep-ocean trenches might be places where the sea floor sank back

FIGURE 2.10 Harry Hess's basic concept of sea-floor spreading. Hess implied, incorrectly, that only the crust moved. We will see that this sketch is an oversimplification.



Take-Home Message

New studies of the sea floor led to the proposal of sea-floor spreading. New sea floor forms at mid-ocean ridges and then moves away from the axis, so ocean basins can get wider with time. Old ocean floor sinks back into the mantle by subduction. As ocean basins grow or shrink, continents drift.

2.5 Evidence for Sea-Floor Spreading

For a hypothesis to become a theory (see Box P.1), researchers must demonstrate that the idea really works. During the 1960s, geologists found that the sea-floor spreading hypothesis successfully explains several previously baffling observations. Here we discuss two: (1) the existence of orderly variations in the strength of the measured magnetic field over the sea floor, producing a pattern of stripes called marine magnetic anomalies; and (2) the variation in sediment thickness on the ocean crust, as measured by drilling.

Marine Magnetic Anomalies

Recognizing anomalies. Geologists can measure the strength of Earth's magnetic field with an instrument called a magnetometer. At any given location on the surface of the Earth, the magnetic field that you measure includes two parts: one produced by the main dipole of the Earth generated by circulation of molten iron in the outer core, and another produced by the magnetism of near-surface rock. A **magnetic anomaly** is the difference between the *expected* strength of the Earth's main dipole field at a certain location and the *actual* measured strength of the magnetic field at that location. Places where the field strength is stronger than expected are *positive* anomalies, and places where the field strength is weaker than expected are *negative* anomalies.

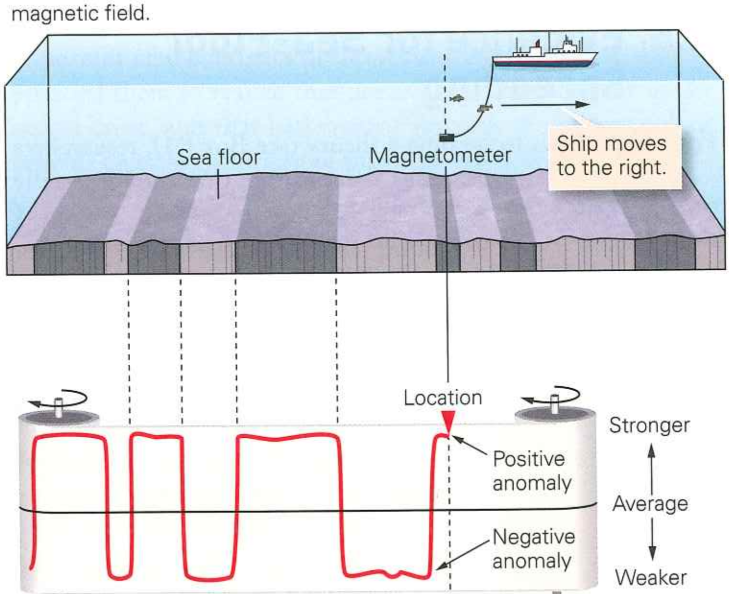
Geologists towed magnetometers back and forth across the ocean to map variations in magnetic field strength (**Fig. 2.11a**). As a ship cruised along its course, the magnetometer's gauge might first detect an interval of strong signal (a positive anomaly) and then an interval of weak signal (a negative anomaly). A graph of signal strength versus distance along the traverse, therefore, has a sawtooth shape (**Fig. 2.11b**). When geologists compiled data from many cruises on a map, these **marine magnetic anomalies** defined distinctive, alternating bands. If we color positive anomalies dark and negative anomalies light, the pattern made by the anomalies resembles the stripes on a candy cane (**Fig. 2.11c**). The mystery of this marine magnetic anomaly pattern, however, remained unsolved until geologists recognized the existence of magnetic reversals.

Magnetic reversals. Recall that Earth's magnetic field can be represented by an arrow, representing the dipole, that presently points from the north magnetic pole to the south magnetic pole. When researchers measured the paleomagnetism of a succession of rock layers that had accumulated over a long period of time, they found that the polarity (which end of a magnet points north and which end points south) of the paleomagnetic field preserved in some layers was the same as that of Earth's present magnetic field, whereas in other layers it was the opposite (**Fig. 2. 12a, b**).

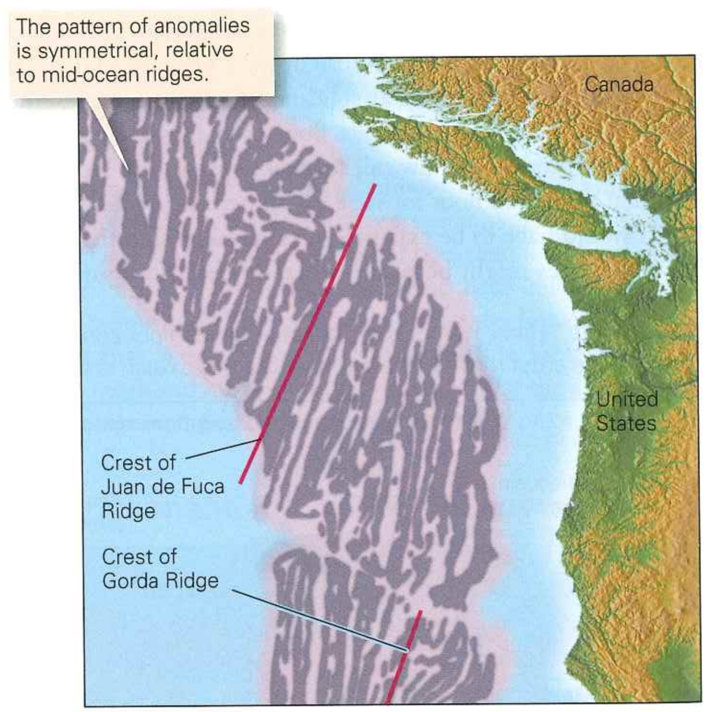
At first, observations of reversed polarity were largely ignored, thought to be the result of lightning strikes or of local chemical reactions between rock and water. But when repeated measurements from around the world revealed a systematic pattern of alternating normal and reversed polarity in rock layers, geologists realized that reversals were a worldwide, not a local, phenomenon. They reached the unavoidable conclusion that, at various times during Earth history, *the polarity of Earth's magnetic field has suddenly reversed*! In other words, sometimes the Earth has normal polarity, as it does today, and sometimes it has reversed polarity (**Fig. 2.12c**). A time when the Earth's field flips from normal to reversed polarity, or vice versa, is called a **magnetic reversal**. When the Earth has reversed polarity, the south magnetic pole lies near the north geographic pole, and the north magnetic pole lies near the south geographic pole. Thus, if you were to use a compass during periods when the Earth's magnetic field was reversed, the north-seeking end of the needle would point to the south

FIGURE 2.11 The discovery of marine magnetic anomalies.

(a) A ship towing a magnetometer detects changes in the strength of the magnetic field.



(b) On a paper record, intervals of stronger magnetism (positive anomalies) alternate with intervals of weaker magnetism (negative anomalies).



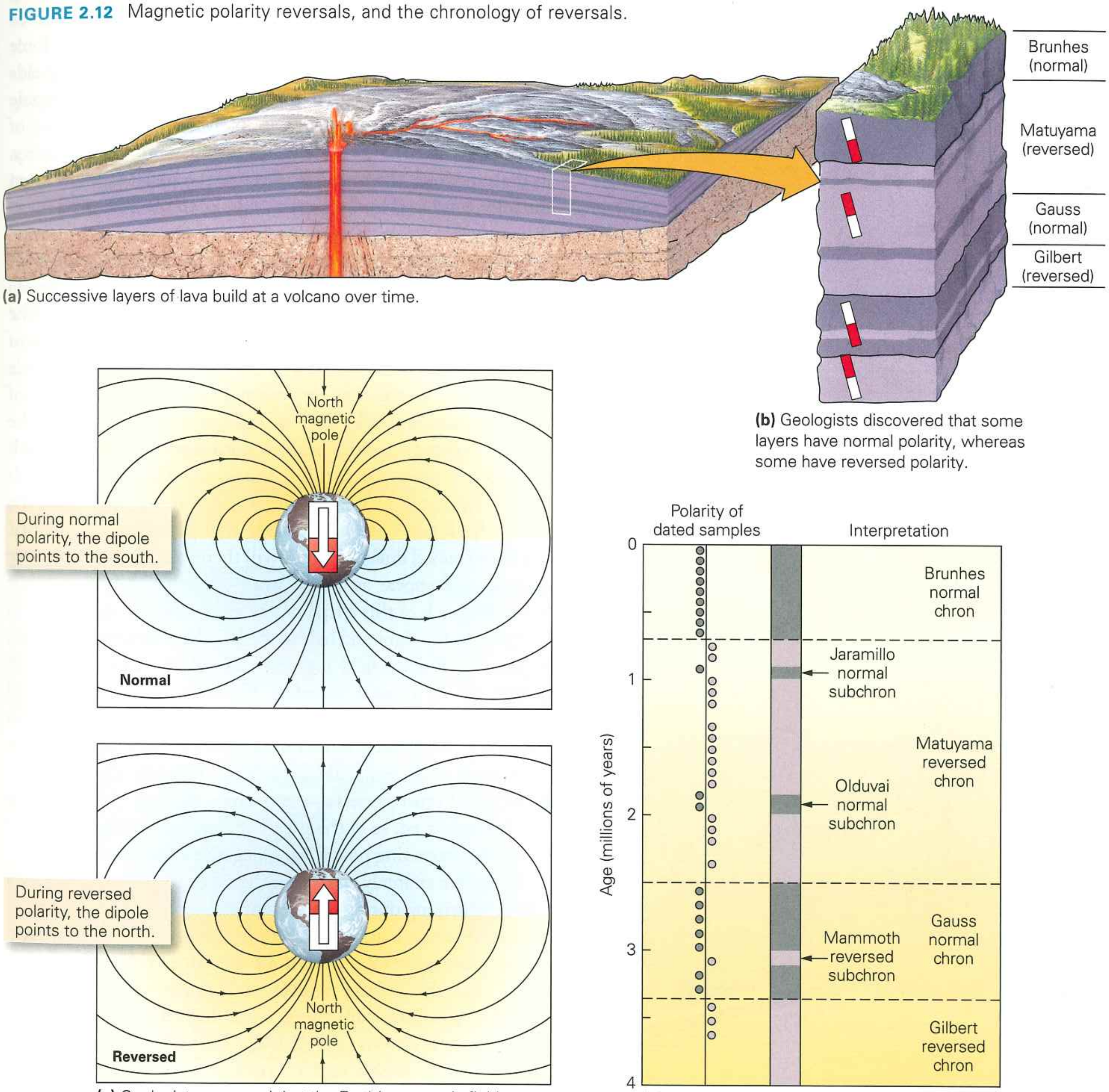
(c) A map showing areas of positive anomalies (dark) and negative anomalies (light) off the west coast of North America. The pattern of anomalies resembles candy-cane strips.

geographic pole. Note that the Earth itself doesn't turn upside down—it is just the magnetic field that reverses.

In the 1950s, about the same time researchers discovered polarity reversals, they developed a technique that permitted them to measure the age of a rock in years. The technique, called isotopic dating, will be discussed in detail in Chapter 10. Geologists applied the technique to determine the ages of rock layers in which they obtained their paleomagnetic measurements, and thus determined *when* the magnetic field of the Earth reversed. With this information, they constructed a history of magnetic reversals for the past 4.5 million years; this history is now called the magnetic-reversal chronology. The time interval between successive reversals is called a **chron**.

A diagram representing the Earth's magnetic-reversal chronology (**Fig. 2.12d**) shows that reversals do not occur regularly, so the lengths of different polarity chrons are different. For example, we have had a normal-polarity chron for about the last 700,000 years. Before that, a reversed-polarity chron occurred. The youngest four polarity chrons (Brunhes, Matuyama, Gauss, and Gilbert) were named after scientists who had made

FIGURE 2.12 Magnetic polarity reversals, and the chronology of reversals.



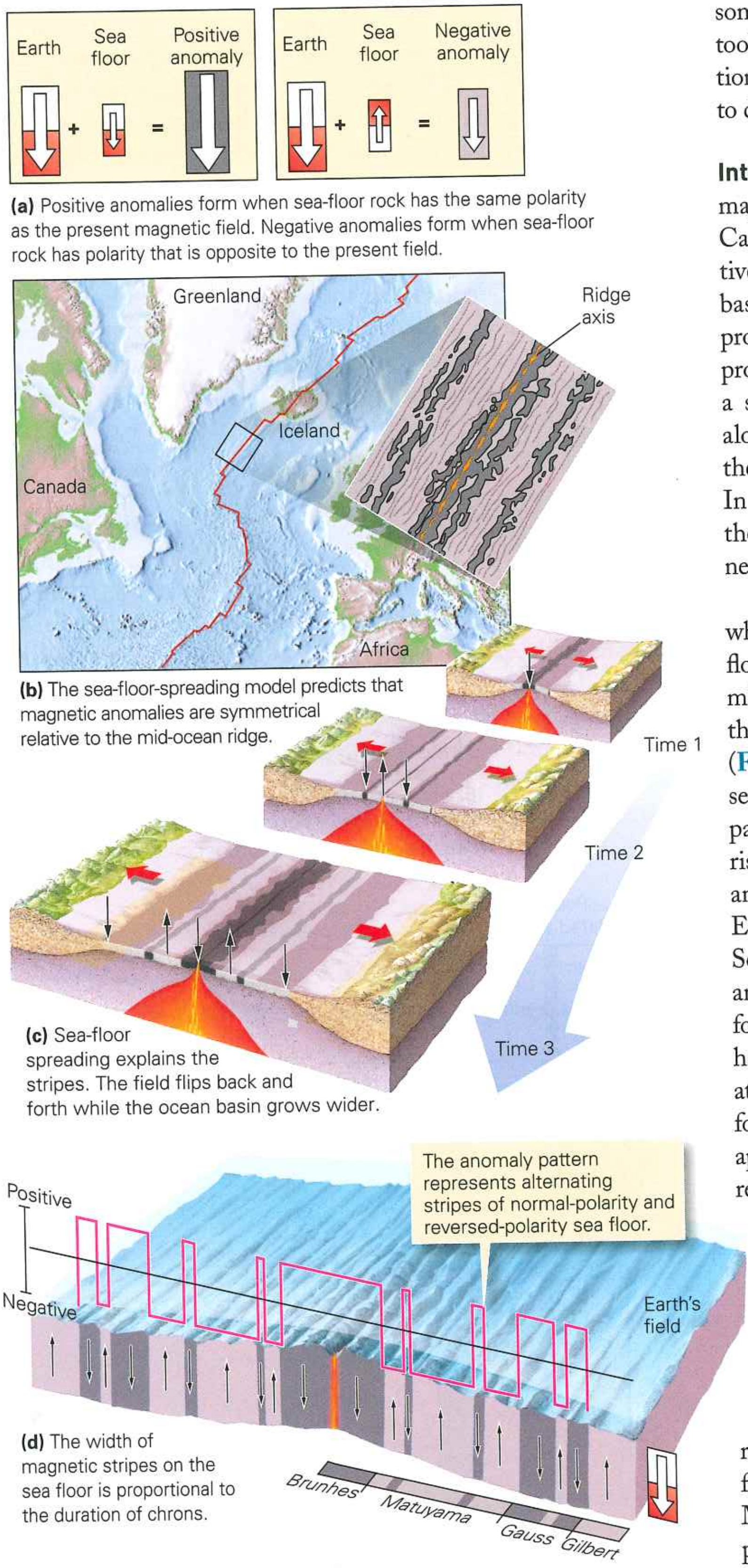
(a) Successive layers of lava build at a volcano over time.

(b) Geologists discovered that some layers have normal polarity, whereas some have reversed polarity.

(c) Geologists proposed that the Earth's magnetic field reverses polarity every now and then.

(d) Observations led to the production of a reversal chronology, with named polarity intervals.

FIGURE 2.13 The progressive development of magnetic anomalies, and the long-term reversal chronology.



(a) Positive anomalies form when sea-floor rock has the same polarity as the present magnetic field. Negative anomalies form when sea-floor rock has polarity that is opposite to the present field.

(b) The sea-floor-spreading model predicts that magnetic anomalies are symmetrical relative to the mid-ocean ridge.

(c) Sea-floor spreading explains the stripes. The field flips back and forth while the ocean basin grows wider.

(d) The width of magnetic stripes on the sea floor is proportional to the duration of chrons.

important contributions to the study of magnetism. As more measurements became available, investigators realized that some short-duration reversals (less than 200,000 years long) took place within the chrons, and they called these shorter durations "polarity subchrons." Using isotopic dating, it was possible to determine the age of chrons back to 4.5 Ma.

Interpreting marine magnetic anomalies. Why do marine magnetic anomalies exist? In 1963, researchers in Britain and Canada proposed a solution to this riddle. Simply put, a positive anomaly occurs over areas of the sea floor where underlying basalt has normal polarity. In these areas, the magnetic force produced by the magnetite grains in basalt *adds* to the force produced by the Earth's dipole—the sum of these forces yields a stronger magnetic signal than expected due to the dipole alone (**Fig. 2.13a**). A negative anomaly occurs over regions of the sea floor where the underlying basalt has a reversed polarity. In these regions, the magnetic force of the basalt *subtracts* from the force produced by the Earth's dipole, so the measured magnetic signal is weaker than expected.

The sea-floor-spreading model easily explains not only why positive and negative magnetic anomalies exist over the sea floor, but also why they define stripes that trend parallel to the mid-ocean ridge and why the pattern of stripes on one side of the ridge is the mirror image of the pattern on the other side (**Fig. 2.13b**). To see why, let's examine stages in the process of sea-floor spreading (**Fig. 2.13c**). Imagine that at Time 1 in the past, the Earth's magnetic field has normal polarity. As the basalt rising at the mid-ocean ridge during this time interval cools and solidifies, the tiny magnetic grains in basalt align with the Earth's field, and thus the rock as a whole has a normal polarity. Sea floor formed during Time 1 will therefore generate a positive anomaly and appear as a dark stripe on an anomaly map. As it forms, the rock of this stripe moves away from the ridge axis, so half goes to the right and half to the left. Now imagine that later, at Time 2, Earth's field has reversed polarity. Sea-floor basalt formed during Time 2, therefore, has reversed polarity and will appear as a light stripe on an anomaly map. As it forms, this reversed-polarity stripe moves away from the ridge axis, and even younger crust forms along the axis. The basalt in each new stripe of crust preserves the polarity that was present at the time it formed, so as the Earth's magnetic field flips back and forth, alternating positive and negative anomaly stripes form. A positive anomaly exists over the ridge axis today because sea floor is forming during the present chron of normal polarity.

Closer examination of a sea-floor magnetic anomaly map reveals that anomalies are not all the same width. Geologists found that the relative widths of anomaly stripes near the Mid-Atlantic Ridge are the same as the relative durations of paleomagnetic chrons (**Fig. 2.13d**). This relationship between anomaly-stripe width and polarity-chron duration indicates that the rate of sea-floor spreading has been constant along the Mid-Atlantic Ridge for at least the last 4.5 million years. If you assume that the spreading rate was constant for tens to hundreds of millions of years, then it is possible to estimate the age of stripes right up to the edge of the ocean.

Evidence from Deep-Sea Drilling

In the late 1960s, a research drilling ship called the *Glomar Challenger* set out to sail around the ocean drilling holes into the sea floor. This amazing ship could lower enough drill pipe to drill in 5-km-deep water and could continue to drill until the hole reached a depth of about 1.7 km (1.1 miles) below the sea floor. Drillers brought up cores of rock and sediment that geoscientists then studied on board.

On one of its early cruises, the *Glomar Challenger* drilled a series of holes through sea-floor sediment to the basalt layer. These holes were spaced at progressively greater distances from the axis of the Mid-Atlantic Ridge. If the model of sea-floor spreading was correct, then not only should the sediment layer be progressively thicker away from the axis, but the age of the oldest sediment just above the basalt should be progressively older away from the axis. When the drilling and the analyses were complete, the prediction was confirmed. Thus, studies of both marine magnetic anomalies and the age of the sea floor proved the sea-floor-spreading model.

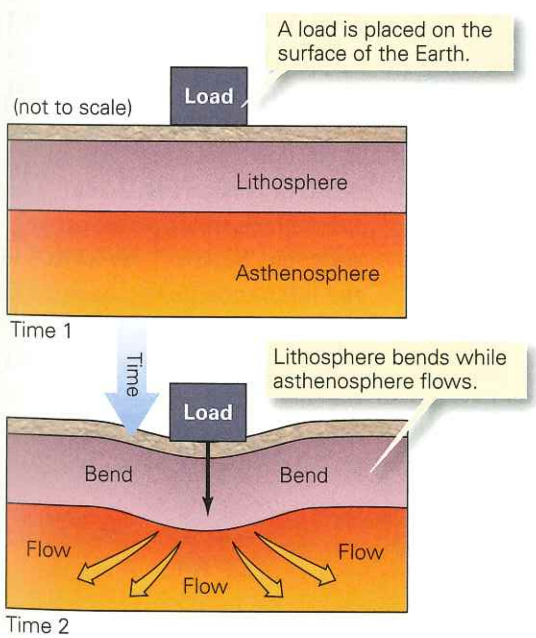
Take-Home Message

Marine magnetic anomalies form because reversals of the Earth's magnetic polarity take place while sea-floor spreading occurs. The discovery of these anomalies, as well as documentation that the sea floor gets older away from the ridge axis, proved that the sea-floor-spreading hypothesis is correct.

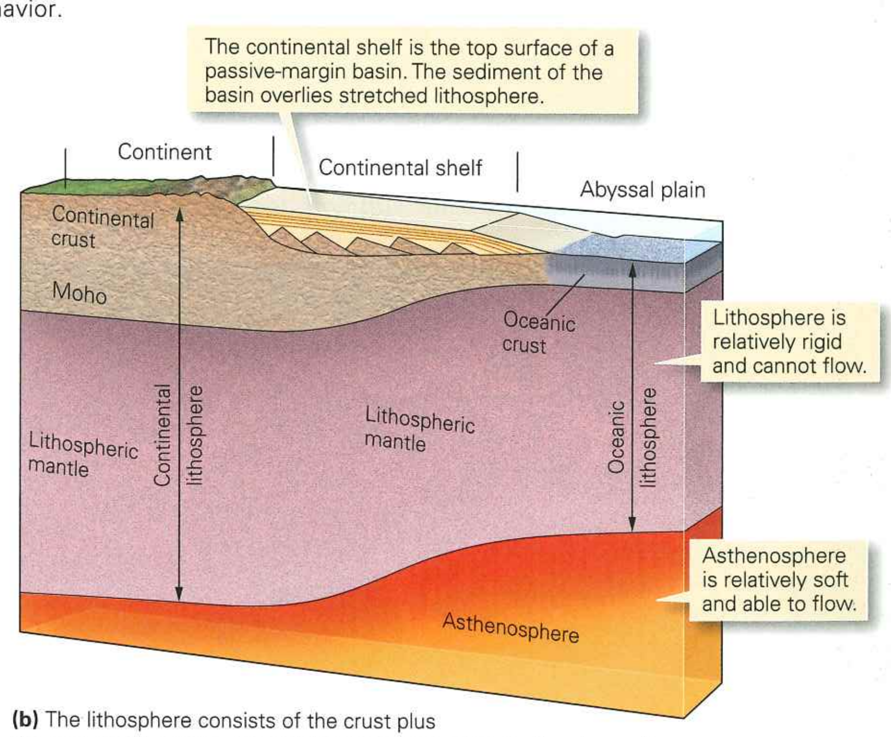
2.6 What Do We Mean by Plate Tectonics?

The paleomagnetic proof of continental drift and the discovery of sea-floor spreading set off a scientific revolution in geology in the 1960s and 1970s. Geologists realized that many of their existing interpretations of global geology, based on the premise that the positions of continents and oceans remain fixed in position through time, were simply wrong! Researchers dropped what they were doing and turned their attention to studying the broader implications of continental drift and sea-floor spreading. It became clear that these phenomena

FIGURE 2.14 Nature of the lithosphere and its behavior.

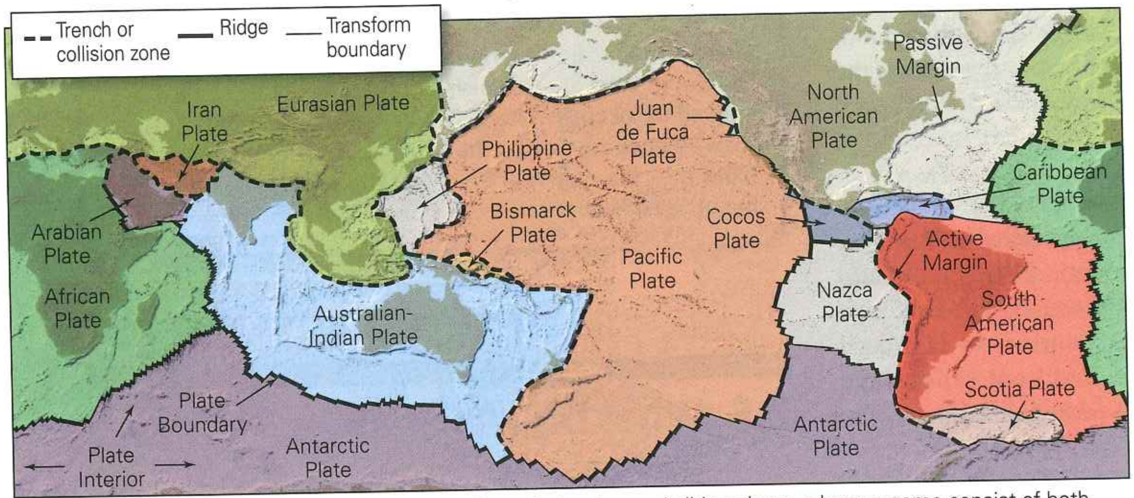


(a) The lithosphere is fairly rigid, but when a heavy load, such as a glacier or volcano, builds on its surface, the surface bends down. This can happen because the underlying "plastic" asthenosphere can flow out of the way.

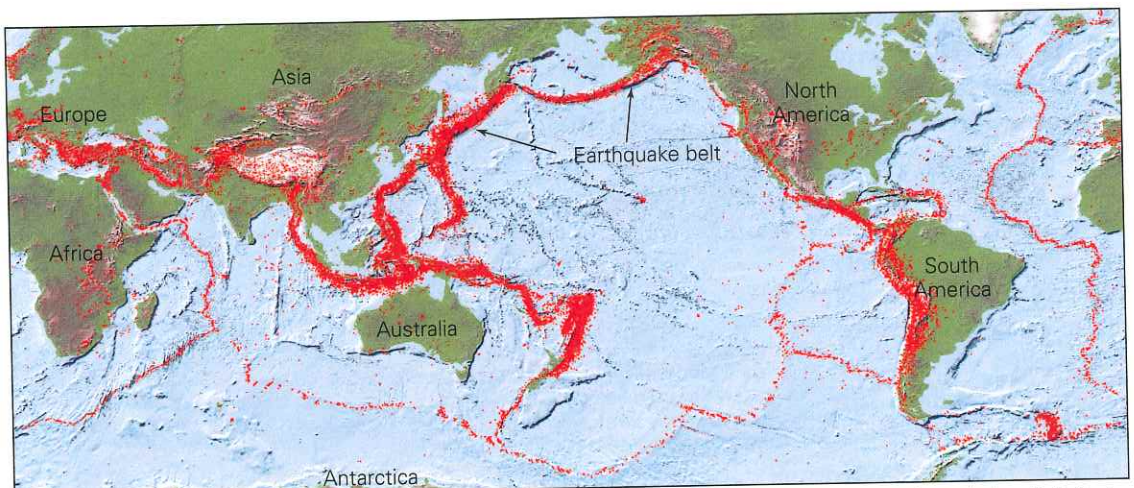


(b) The lithosphere consists of the crust plus the uppermost mantle. It is thicker beneath continents than beneath oceans.

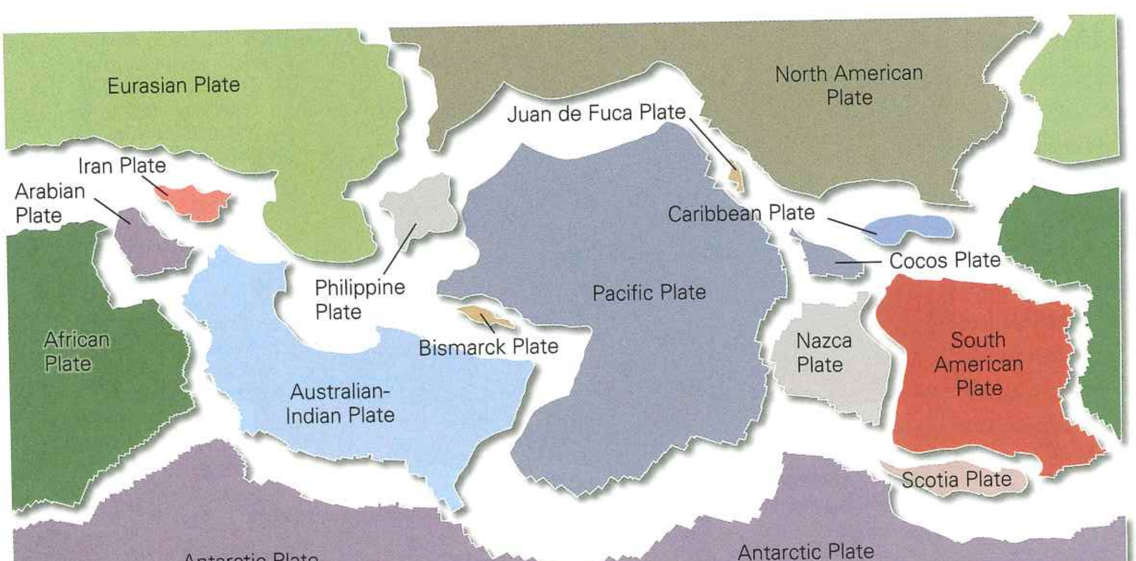
FIGURE 2.15 The locations of plate boundaries and the distribution of earthquakes.



(a) A map of major plates shows that some consist entirely of oceanic lithosphere, whereas some consist of both continental and oceanic lithosphere. Active continental margins lie along plate boundaries; passive margins do not.



(b) The locations of earthquakes (red dots) mostly fall in distinct bands that correspond to plate boundaries. Relatively few earthquakes occur in the stabler plate interiors.



(c) An exploded view of the plates emphasizes the variation in shape and size of the plates.

required that the outer shell of the Earth was divided into rigid plates that moved relative to each other. New studies clarified the meaning of a plate, defined the types of plate boundaries, constrained plate motions, related plate motions to earthquakes and volcanoes, showed how plate interactions can explain mountain belts and seamount chains, and outlined the history of past plate motions. From these, the modern theory of plate tectonics evolved. Below, we first describe lithosphere plates and their boundaries, and then outline the basic principles of plate tectonics theory.

The Concept of a Lithosphere Plate

We learned earlier that geoscientists divide the outer part of the Earth into two layers. The **lithosphere** consists of the crust plus the top (cooler) part of the upper mantle. It behaves relatively rigidly, meaning that when a force pushes or pulls on it, it does not flow but rather bends or breaks (**Fig. 2.14a**). The lithosphere floats on a relatively soft, or "plastic," layer called the **asthenosphere**, composed of warmer (> 1280°C) mantle that can flow slowly when acted on by a force. As a result, the asthenosphere convects, like water in a pot, though much more slowly.

Continental lithosphere and oceanic lithosphere differ markedly in their thicknesses. On average, continental lithosphere has a thickness of 150 km, whereas old oceanic lithosphere has a thickness of about 100 km (**Fig. 2.14b**).

(For reasons discussed later in this chapter, new oceanic lithosphere at a mid-ocean ridge is much thinner.) Recall that the crustal part of continental lithosphere ranges from 25 to 70 km thick and consists largely of low-density felsic and intermediate rock (see Chapter 1). In contrast, the crustal part of oceanic lithosphere is only 7 to 10 km thick and consists largely of relatively high-density mafic rock (basalt and gabbro). The mantle part of both continental and oceanic lithosphere consists of very high-density ultramafic rock (peridotite). Because of these differences, the continental lithosphere "floats" at a higher level than does the oceanic lithosphere.

The lithosphere forms the Earth's relatively rigid shell. But unlike the shell of a hen's egg, the lithospheric shell contains a number of major breaks, which separate it into distinct pieces. As noted earlier, we call the pieces **lithosphere plates**, or simply plates. The breaks between plates are known as **plate boundaries** (**Fig. 2.15a**). Geoscientists distinguish twelve major plates and several microplates.

The Basic Principles of Plate Tectonics

With the background provided above, we can restate plate tectonics theory concisely as follows. The Earth's lithosphere is divided into plates that move relative to each other. As a plate moves, its internal area remains mostly, but not perfectly, rigid and intact. But rock along plate boundaries undergoes intense deformation (cracking, sliding, bending, stretching, and squashing) as the plate grinds or scrapes against its neighbors or pulls away from its neighbors. As plates move, so do the continents that form part of the plates. Because of plate tectonics, the map of Earth's surface constantly changes.

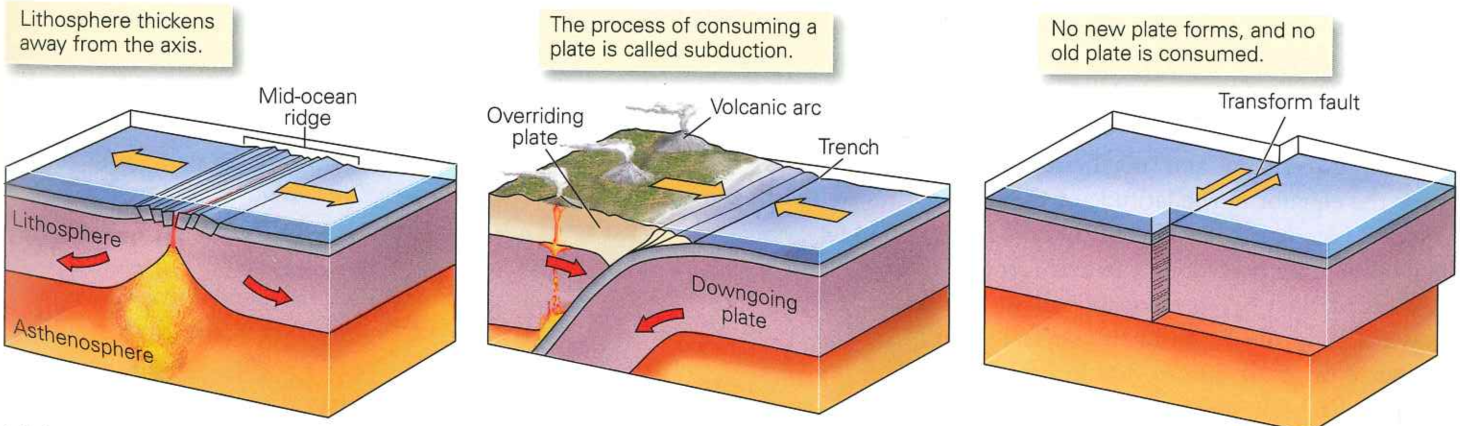
Identifying Plate Boundaries

How do we recognize the location of a plate boundary? The answer becomes clear from looking at a map showing the locations of earthquakes (**Fig. 2.15b**). Recall from Chapter 1 that earthquakes are vibrations caused by shock waves that are generated where rock breaks and suddenly slips along a fault. The epicenter marks the point on the Earth's surface directly above the earthquake. Earthquake epicenters do not speckle the globe randomly, like buckshot on a target. Rather, the majority occur in relatively narrow, distinct belts. These earthquake belts define the position of plate boundaries because the fracturing and slipping that occurs along plate boundaries generates earthquakes. Plate interiors, regions away from the plate boundaries, remain relatively earthquake-free because they do not accommodate as much movement. While earthquakes serve as the most definitive indicator of a plate boundary, other prominent geologic features also develop along plate boundaries, as you will learn by the end of this chapter.

Note that some plates consist entirely of oceanic lithosphere, whereas some plates consist of both oceanic and continental lithosphere. Also, note that not all plates are the same size (**Fig. 2.15c**). Some plate boundaries follow continental margins, the boundary between a continent and an ocean, but others do not. For this reason, we distinguish between **active margins**, which are plate boundaries, and **passive margins**, which are not plate boundaries. Earthquakes are common at active margins, but not at passive margins. Along passive margins, continental crust is thinner than in continental interiors. Thick (10 to 15 km) accumulations of sediment cover this thinned crust. The surface of this sediment

Did you ever wonder . . .
why earthquakes don't occur everywhere?

FIGURE 2.16 The three types of plate boundaries differ based on the nature of relative movement.



(a) At a divergent boundary, two plates move away from the axis of a mid-ocean ridge. New oceanic lithosphere forms.

(b) At a convergent boundary, two plates move toward each other; the downgoing plate sinks beneath the overriding plate.

(c) At a transform boundary, two plates slide past each other on a vertical fault surface.

layer is a broad, shallow (less than 500 m deep) region called the **continental shelf**, home to the major fisheries of the world.

Geologists define three types of plate boundaries, based simply on the relative motions of the plates on either side of the boundary (**Fig. 2.16a–c**). A boundary at which two plates move apart from each other is a **divergent boundary**. A boundary at which two plates move toward each other so that one plate sinks beneath the other is a **convergent boundary**. And a boundary at which two plates slide sideways past each other is a **transform boundary**. Each type of boundary looks and behaves differently from the others, as we will now see.

Take-Home Message

Earth's lithosphere is divided into about twenty plates that move relative to each other. Geologists recognize three different types of plate boundaries (divergent, convergent, and transform) based on relative motion across the boundary. Plate boundaries are defined by seismic belts.

2.7 Divergent Plate Boundaries and Sea-Floor Spreading

At a divergent boundary, or spreading boundary, two oceanic plates move apart by the process of sea-floor spreading. Note that an *open space does not develop* between diverging plates. Rather, as the plates move apart, new oceanic lithosphere forms continually along the divergent boundary (**Fig. 2.17a**). This process takes place at a submarine mountain range called a mid-ocean ridge that rises 2 km above the adjacent abyssal plains of the ocean. Thus, geologists commonly refer to a divergent boundary as a mid-ocean ridge, or simply a ridge. Water depth above ridges averages about 2.5 km.

To characterize a divergent boundary more completely, let's look at one mid-ocean ridge in more detail (**Fig. 2.17b**). The Mid-Atlantic Ridge extends from the waters between northern Greenland and northern Scandinavia southward across the equator to the latitude of the southern tip of South America. Geologists have found that the formation of new sea floor takes place *only* along the axis (centerline) of the ridge, which is marked by an elongate valley. The sea floor slopes away, reaching the depth of the abyssal plain (4 to 5 km) at a distance of about 500 to 800 km from the ridge axis (see Fig. 2.7). Roughly speaking, the Mid-Atlantic Ridge is symmetrical—its eastern half looks like a mirror image of its western half. The ridge consists, along its length, of short segments (tens to hundreds of km long) that step over at breaks that, as we noted earlier, are called fracture zones. Later, we will see that these correspond to transform faults.

How Does Oceanic Crust Form at a Mid-Ocean Ridge?

As sea-floor spreading takes place, hot asthenosphere rises beneath the ridge and begins to melt, and molten rock, or magma, forms (**Fig. 2.17c**). We will explain why this magma forms in Chapter 4. Magma has a lower density than solid rock, so it behaves buoyantly and rises, as oil rises above vinegar in salad dressing. Molten rock eventually accumulates in the crust below the ridge axis, filling a region called a magma chamber. As the magma cools, it turns into a mush of crystals. Some of the magma solidifies completely along the side of the chamber to make the coarse-grained, mafic igneous rock called gabbro. The rest rises still higher to fill vertical cracks, where it solidifies and forms wall-like sheets, or dikes, of basalt. Some magma rises all the way to the surface of the sea floor at the ridge axis and spills out of small submarine volcanoes. The resulting lava cools to form a layer of basalt blobs called pillows. Observers in research submarines have detected chimneys spewing hot, mineralized water rising from cracks in the sea floor along the ridge axis. These chimneys are called **black smokers** because the water they emit looks like a cloud of dark smoke; the color comes from a suspension of tiny mineral grains that precipitate in the water the instant that the water cools (**Fig. 2.18**).

As soon as it forms, new oceanic crust moves away from the ridge axis, and when this happens, more magma rises from below, so still more crust forms. In other words, like a vast, continuously moving conveyor belt, magma from the mantle rises to the Earth's surface at the ridge, solidifies to form oceanic crust, and then moves laterally away from the ridge. Because all sea floor forms at mid-ocean ridges, the youngest sea floor occurs on either side of the ridge axis, and sea floor becomes progressively older away from the ridge. In the Atlantic Ocean, the oldest sea floor, therefore, lies adjacent to the passive continental margins on either side of the ocean (**Fig. 2.19**). The oldest ocean floor on our planet underlies the western Pacific Ocean; this crust formed about 200 million years ago.

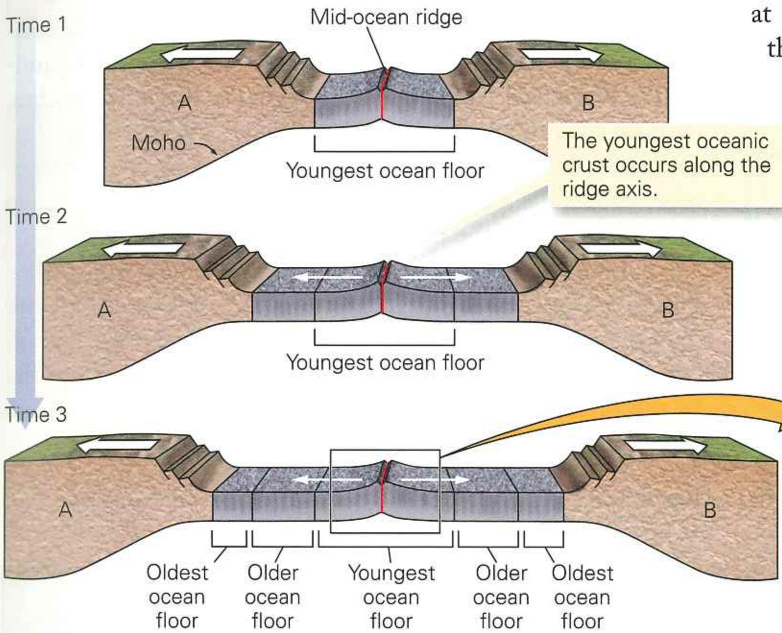
The tension (stretching force) applied to newly formed solid crust as spreading takes place breaks the crust, resulting in the formation of faults. Slip on the faults causes divergent-boundary earthquakes and produces numerous cliffs, or scarps, that lie parallel to the ridge axis.

How Does the Lithospheric Mantle Form at a Mid-Ocean Ridge?

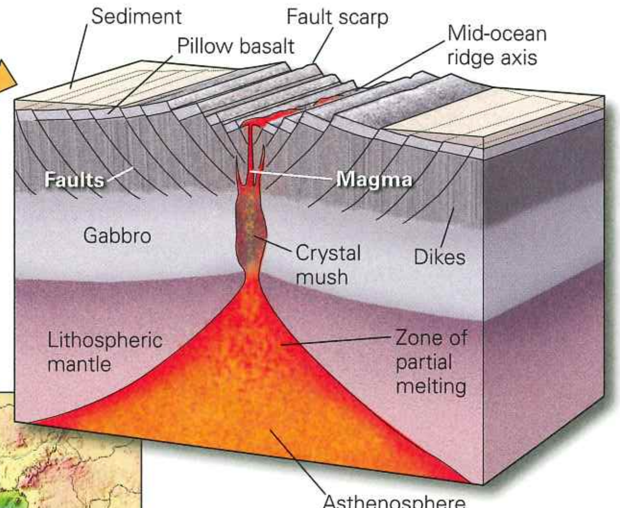
So far, we've seen how oceanic *crust* forms at mid-ocean ridges. How does the *mantle* part of the oceanic lithosphere form? This part consists of the cooler uppermost layer of the mantle, in which temperatures are less than about 1,280°C. At the ridge axis, such temperatures occur almost at the base of the crust, because of the presence of rising hot asthenosphere and hot magma, so lithospheric mantle beneath the ridge axis effectively doesn't exist. But as the newly formed oceanic crust moves away from the ridge axis, the crust and the uppermost mantle directly beneath it gradually cool by losing heat to the ocean above. As soon as mantle rock cools below 1,280°C, it becomes, by definition, part of the lithosphere.

As oceanic lithosphere continues to move away from the ridge axis, it continues to cool, so the lithospheric mantle, and therefore the oceanic lithosphere as a whole, grows progressively thicker (**Fig. 2.20a, b**). Note that this process doesn't change the thickness of the oceanic crust, for the crust formed entirely at the ridge axis. The rate at which cooling and lithospheric thickening occur decreases progressively with increasing distance from the ridge axis. In fact, by the time the lithosphere is about 80 million years old, it has just about reached its maximum thickness. As lithosphere thickens and gets cooler and denser, it sinks down into the asthenosphere, like a ship taking on ballast. Thus, the ocean is deeper over older ocean floor than over younger ocean floor.

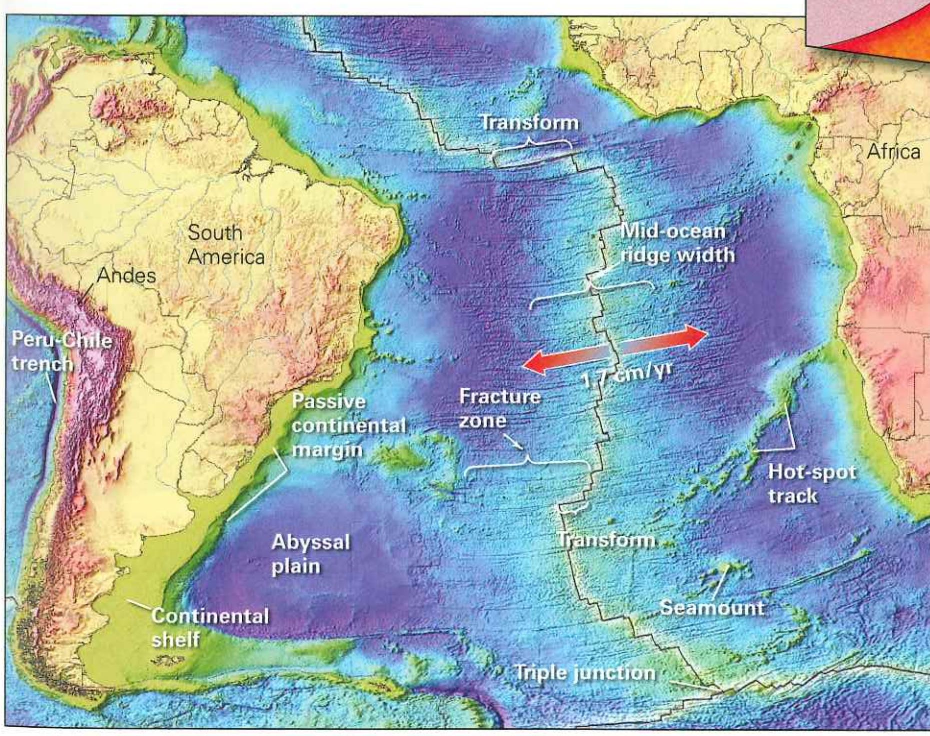
FIGURE 2.17 The process of sea-floor spreading.



(a) As sea-floor spreading progresses, new oceanic lithosphere forms at the mid-ocean ridge axis. For simplicity, only the crust is shown.



(c) Architecture of a mid-ocean ridge, the site of sea-floor spreading. Some magma freezes into new rock within the crust, whereas some spills out onto the surface of the sea floor. Faults break up the crust as it stretches apart.



(b) The bathymetry of the Mid-Atlantic Ridge in the Southern Atlantic Ocean. The lighter shades of blue are shallower depths.

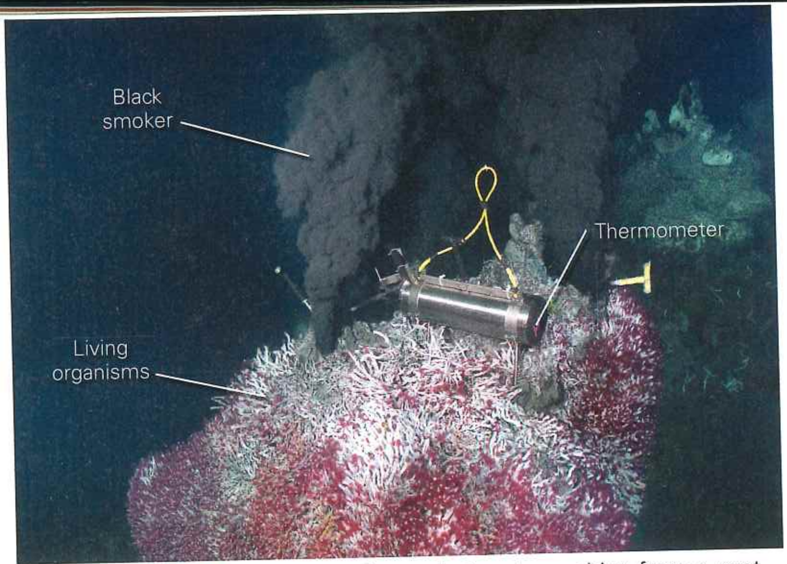


FIGURE 2.18 A column of superhot water gushing from a vent known as a black smoker along the mid-ocean ridge. A local ecosystem of bacteria, shrimp, and worms lives around the vent.

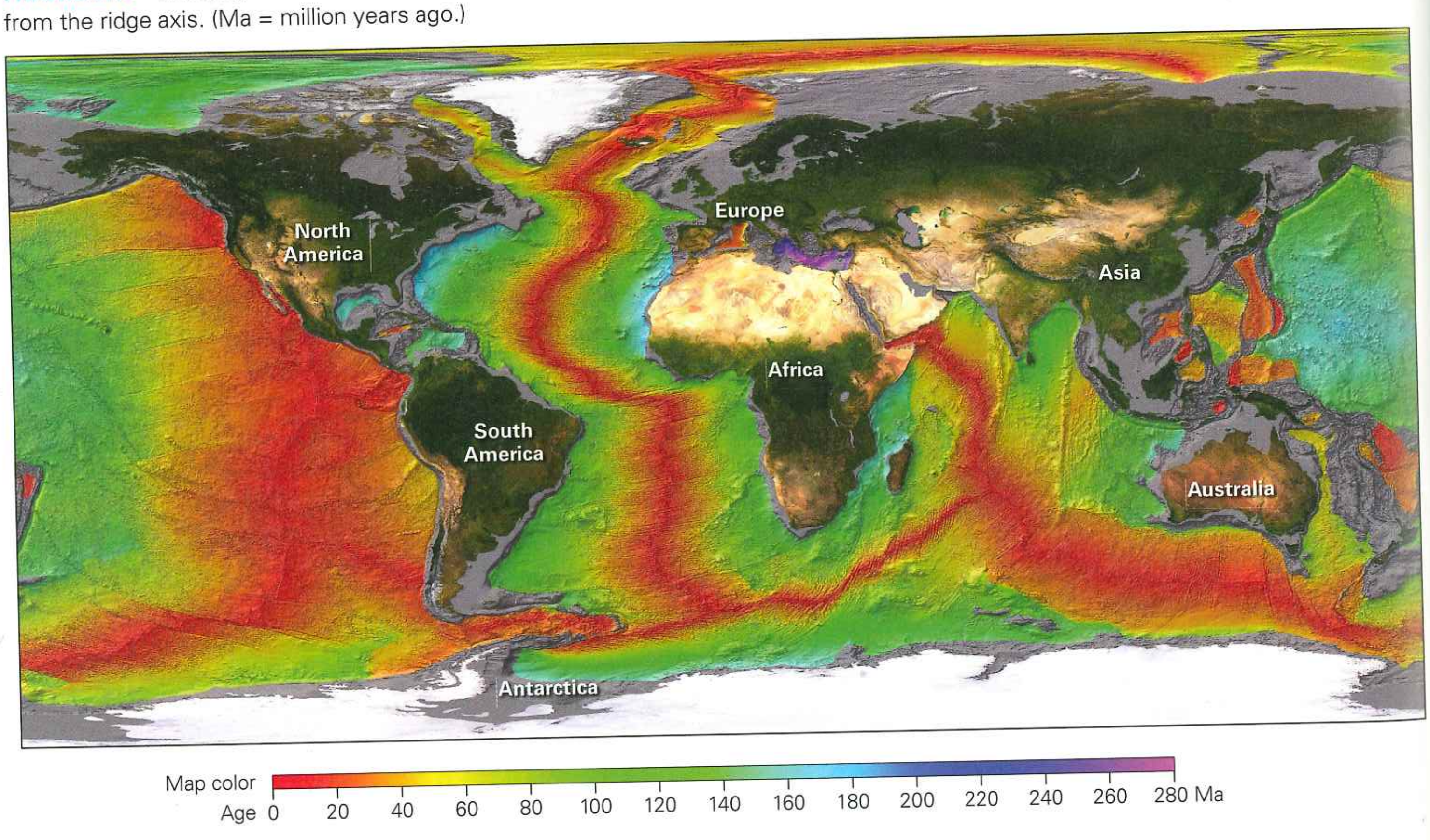
> **Take-Home Message**
>
> **Sea-floor spreading occurs at divergent plate boundaries, defined by mid-ocean ridges. New oceanic crust solidifies from basaltic magma along the ridge axis. As plates move away from the axis, they cool, and the lithospheric mantle forms and thickens.**

2.8 Convergent Plate Boundaries and Subduction

At convergent plate boundaries, two plates, at least one of which is oceanic, move toward one another. But rather than butting each other like angry rams, one oceanic plate bends and sinks down into the asthenosphere beneath the other plate. Geologists refer to the sinking process as **subduction**, so convergent boundaries are also known as subduction zones. Because subduction at a convergent boundary consumes old ocean lithosphere and thus "consumes" oceanic basins, geologists also refer to convergent boundaries as consuming boundaries, and because they are delineated by deep-ocean trenches, they are sometimes simply called **trenches** (see Fig. 2.8). The amount of oceanic plate consumption worldwide, averaged over time, equals the amount of sea-floor spreading worldwide, so the surface area of the Earth remains constant through time.

Subduction occurs for a simple reason: oceanic lithosphere, once it has aged at least 10 million years, is denser than the underlying asthenosphere and thus can sink through the asthenosphere if given an opportunity. Where it lies flat on the surface of the asthenosphere, oceanic lithosphere can't

FIGURE 2.19 This map of the world shows the age of the sea floor. Note how the sea floor grows older with increasing distance from the ridge axis. (Ma = million years ago.)



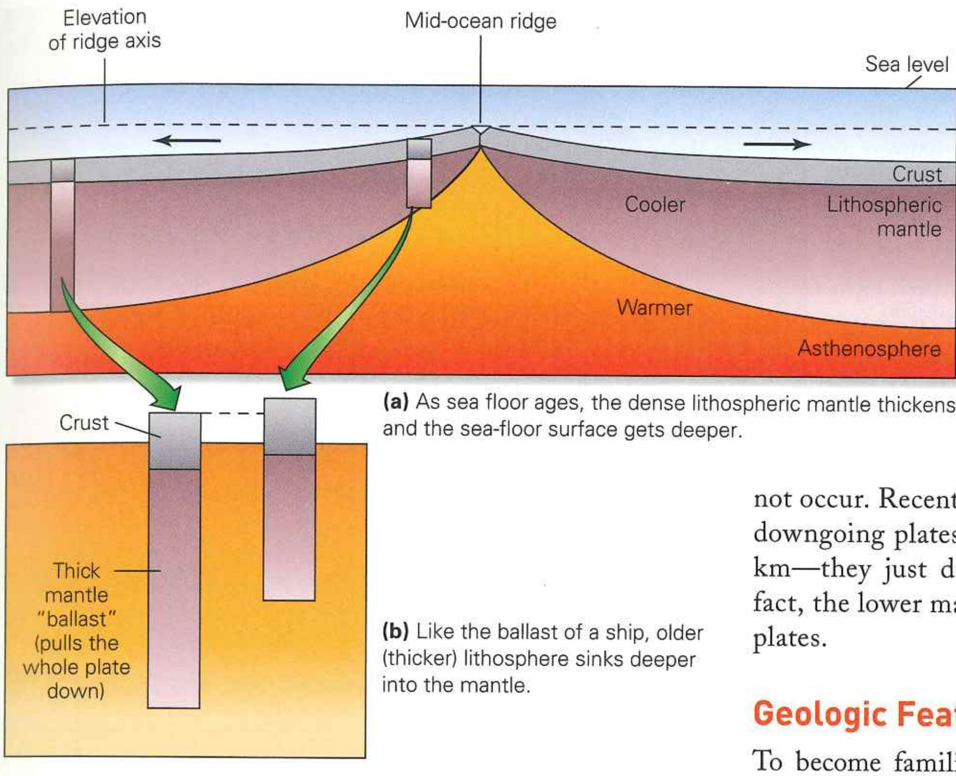


FIGURE 2.20 Changes accompanying the aging of lithosphere.

(a) As sea floor ages, the dense lithospheric mantle thickens and the sea-floor surface gets deeper.

(b) Like the ballast of a ship, older (thicker) lithosphere sinks deeper into the mantle.

sink. However, once the end of the convergent plate bends down and slips into the mantle, it continues downward like an anchor falling to the bottom of a lake (**Fig. 2.21a**). As the lithosphere sinks, asthenosphere flows out of its way, just as water flows out of the way of a sinking anchor. But unlike water, the asthenosphere can flow only very slowly, so oceanic lithosphere can sink only very slowly, at a rate of less than about 15 cm per year. To visualize the difference, imagine how much faster a coin can sink through water than it can through honey.

Note that the "downgoing plate," the plate that has been subducted, *must* be composed of oceanic lithosphere. The overriding plate, which does not sink, can consist of either oceanic or continental lithosphere. Continental crust cannot be subducted because it is too buoyant; the low-density rocks of continental crust act like a life preserver keeping the continent afloat. If continental crust moves into a convergent margin, subduction eventually stops. Because of subduction, all ocean floor on the planet is less than about 200 million years old. Because continental crust cannot subduct, some continental crust has persisted at the surface of the Earth for over 3.8 billion years.

Earthquakes and the Fate of Subducted Plates

At convergent plate boundaries, the downgoing plate grinds along the base of the overriding plate, a process that generates large earthquakes. These earthquakes occur fairly close to the Earth's surface, so some of them cause massive destruction in coastal cities. But earthquakes also happen in downgoing plates at greater depths. In fact, geologists have detected earthquakes within downgoing plates to a depth of 660 km. The band of earthquakes in a downgoing plate is called a **Wadati-Benioff zone**, after its two discoverers (**Fig. 2.21b**).

At depths greater than 660 km, conditions leading to earthquakes in subducted lithosphere evidently do not occur. Recent observations, however, indicate that some downgoing plates do continue to sink *below* a depth of 660 km—they just do so without generating earthquakes. In fact, the lower mantle may be a graveyard for old subducted plates.

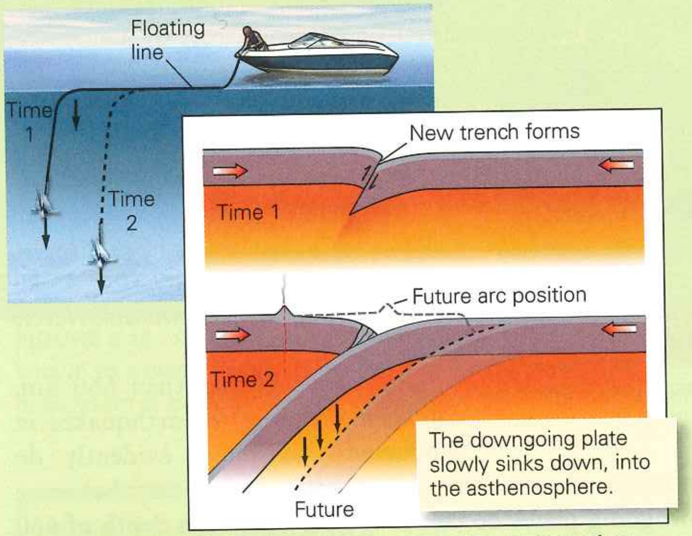
Geologic Features of a Convergent Boundary

To become familiar with the various geologic features that occur along a convergent plate boundary, let's look at an example, the boundary between the western coast of the South American Plate and the eastern edge of the Nazca Plate (a portion of the Pacific Ocean floor). A deep-ocean trench, the Peru-Chile Trench, delineates this boundary (see Fig. 2.17b). Such trenches form where the plate bends as it starts to sink into the asthenosphere.

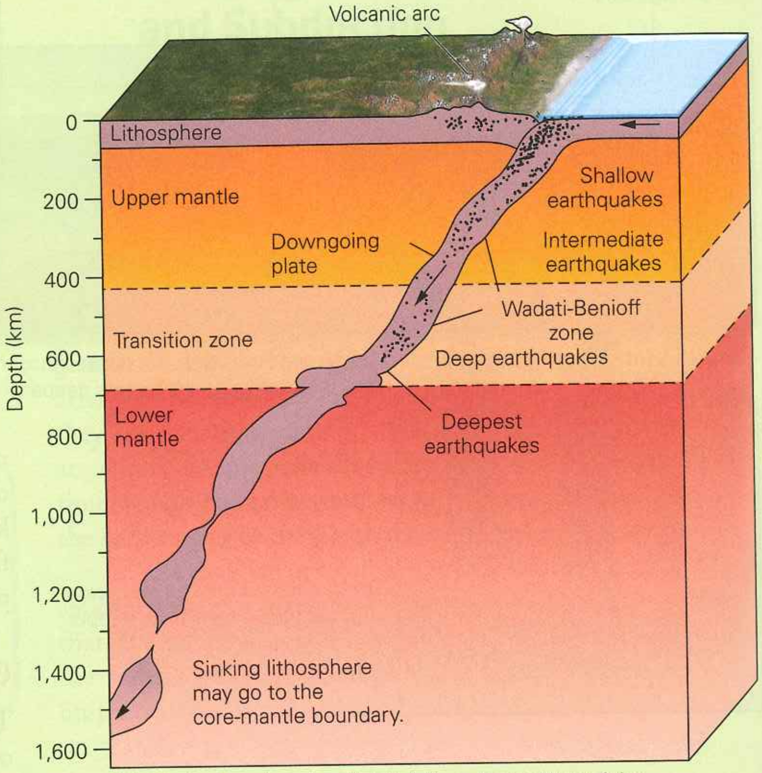
In the Peru-Chile Trench, as the downgoing plate slides under the overriding plate, sediment (clay and plankton) that had settled on the surface of the downgoing plate, as well as sand that fell into the trench from the shores of South America, gets scraped up and incorporated in a wedge-shaped mass known as an **accretionary prism** (**Fig. 2.21c**). An accretionary prism forms in basically the same way as a pile of snow or sand in front of a plow, and like snow, the sediment tends to be squashed and contorted.

A chain of volcanoes known as a **volcanic arc** develops behind the accretionary prism (**See for Yourself B**). As we will see in Chapter 4, the magma that feeds these volcanoes forms just above the surface of the downgoing plate where the plate reaches a depth of about 150 km below the Earth's surface. If the volcanic arc forms where an oceanic plate subducts beneath continental lithosphere, the resulting chain of volcanoes grows on the continent and forms a continental volcanic arc. (In some cases, the plates squeeze together across a continental arc, causing a belt of faults to form behind the arc.) If, however, the volcanic arc grows

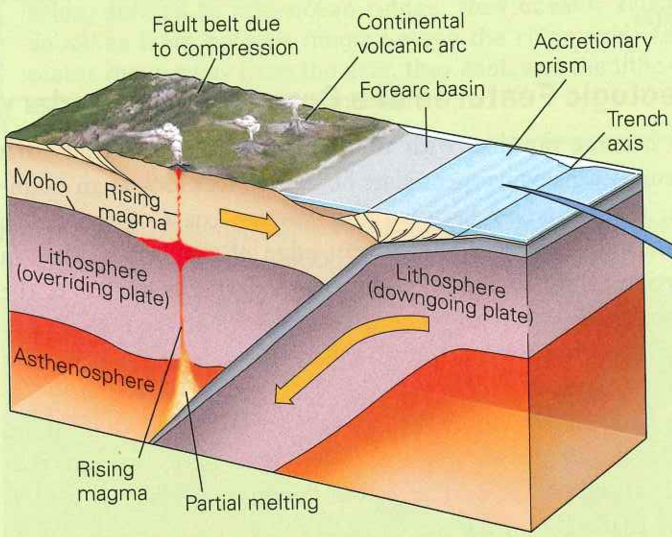
FIGURE 2.21 During the process of subduction, oceanic lithosphere sinks back into the deeper mantle.



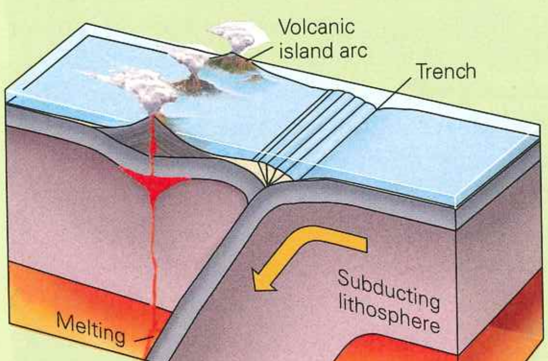
(a) Sinking of the downgoing plate resembles the sinking of an anchor attached to a buoyant rope.



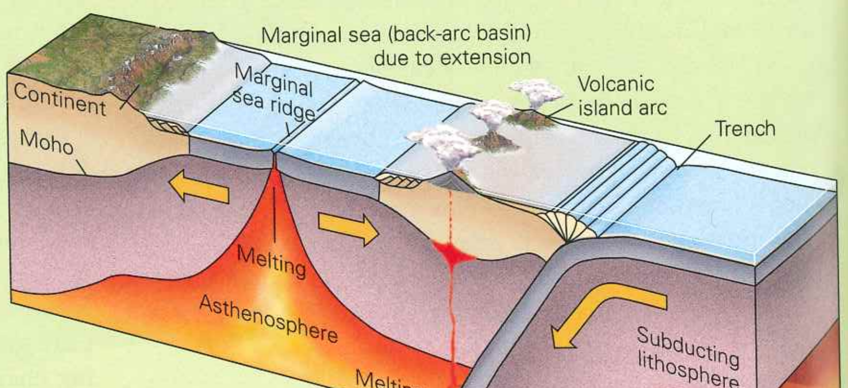
(b) A belt of earthquakes (dots) defines the position of the downgoing plate in the region above a depth of about 660 km.



(c) A model of a subduction zone along the edge of a continent. Note the position of the trench, accretionary prism, and volcanic arc.



(d) A model of a subduction zone along an island arc. Note the position of the back-arc basin, here formed by spreading.



(e) In some cases, an arc may grow on a small piece of continental crust that split away from the main continent when a back-arc basin formed.

where one oceanic plate subducts beneath another oceanic plate, the resulting volcanoes form a chain of islands known as a volcanic island arc (**Fig. 2.21d**). A back-arc basin exists either where subduction happens to begin offshore, trapping ocean lithosphere behind the arc, or where stretching of the lithosphere behind the arc leads to the formation of a small spreading ridge behind the arc (**Fig. 2.21e**).

Take-Home Message

At a convergent plate boundary, an oceanic plate sinks into the mantle beneath the edge of another plate. A volcanic arc and a trench delineate such plate boundaries, and earthquakes happen along the contact between the two plates as well as in the downgoing slab.

2.9 Transform Plate Boundaries

When researchers began to explore the bathymetry of mid-ocean ridges in detail, they discovered that mid-ocean ridges are not long, uninterrupted lines, but rather consist of short segments that appear to be offset laterally from each other (**Fig. 2.22a**) by narrow belts of broken and irregular sea floor. These belts, or **fracture zones**, lie roughly at right angles to the ridge segments, intersect the ends of the segments, and extend beyond the ends of the segments. Originally, researchers incorrectly assumed that the entire length of each fracture zone was a fault, and that slip on a fracture zone had displaced segments of the mid-ocean ridge sideways, relative to each other. In other words, they imagined that a mid-ocean ridge initiated as a continuous, fence-like line that only later was broken up by faulting. But when information about the distribution of earthquakes along mid-ocean ridges became available, it was clear that this model could not be correct. Earthquakes, and therefore active fault slip, occur *only* on the segment of a fracture zone that lies between two ridge segments. The portions of fracture zones that extend beyond the edges of ridge segments, out into the abyssal plain, are not seismically active.

The distribution of movement along fracture zones remained a mystery until a Canadian researcher, J. Tuzo Wilson, began to think about fracture zones in the context of the sea-floor-spreading concept. Wilson proposed that fracture zones formed *at the same time* as the ridge axis itself, and thus the ridge consisted of separate segments to start with. These segments were linked (not offset) by fracture zones. With this idea in mind, he drew a sketch map showing two ridge-axis segments linked by a fracture zone, and he drew arrows to indicate the direction that ocean floor was moving, relative to the ridge axis, as a result of sea-floor-spreading (**Fig. 2.22b**). Look at the arrows in Figure 2.22b. Clearly, the movement direction on the active portion of the fracture zone must be opposite to the movement direction that researchers originally thought occurred on the structure. Further, in Wilson's model, slip occurs only along the segment of the fracture zone between the two ridge segments (**Fig. 2.22c**). Plates on opposite sides of the inactive part of a fracture zone move together, as one plate.

Wilson introduced the term **transform boundary**, or transform fault, for the actively slipping segment of a fracture zone between two ridge segments, and he pointed out that these are a third type of plate boundary. At a transform boundary, one plate slides sideways past another, but no new plate forms and no old plate is consumed. Transform boundaries are, therefore, defined by a vertical fault on which the slip direction parallels the Earth's surface. The slip breaks up the crust and forms a set of steep fractures.

So far we've discussed only transforms along mid-ocean ridges. Not all transforms link ridge segments. Some, such as the Alpine Fault of New Zealand, link trenches, while others link a trench to a ridge segment. Further, not all transform faults occur in oceanic lithosphere; a few cut across continental lithosphere. The San Andreas Fault, for example, which cuts across California, defines part of the plate boundary between the North American Plate and the Pacific Plate—the portion of California that lies to the west of the fault (including Los Angeles) is part of the Pacific Plate, while the portion that lies to the east of the fault is part of the North American Plate (**Fig. 2.22d, e**).

Take-Home Message

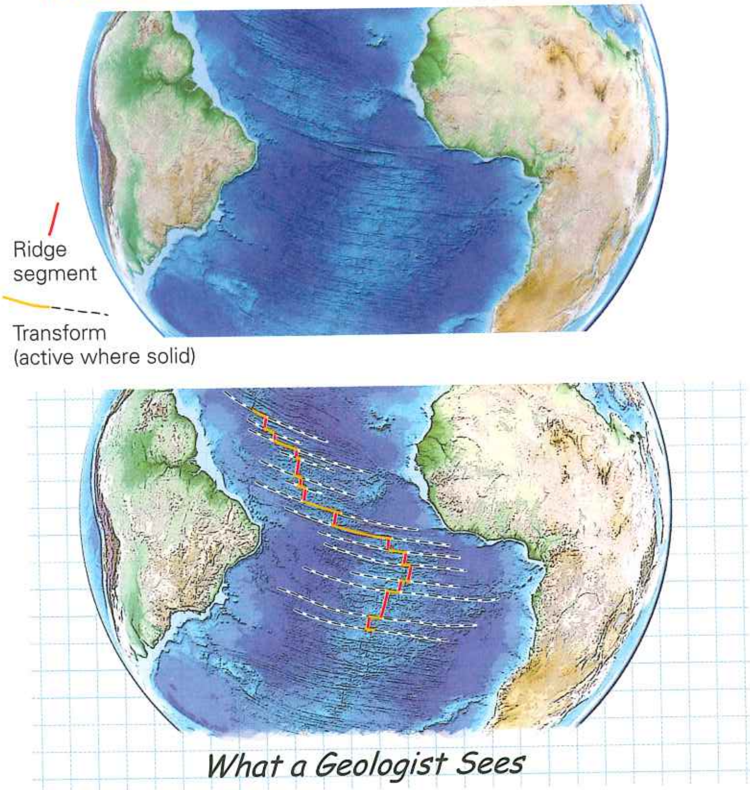
At transform plate boundaries, one plate slips sideways past another. Most transform boundaries link segments of mid-ocean ridges, but some, such as the San Andreas Fault, cut across continental crust.

2.10 Special Locations in the Plate Mosaic

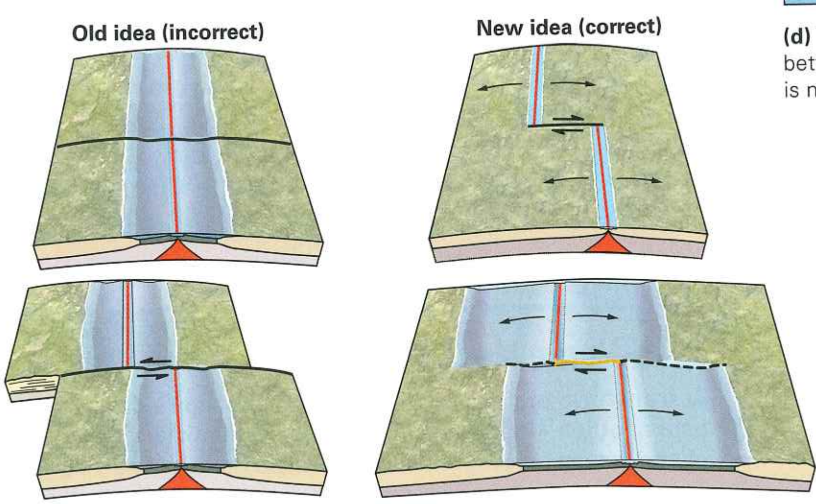
Triple Junctions

Geologists refer to a place where three plate boundaries intersect as a **triple junction**, and name them after the types of boundaries that intersect. For example, the triple junction formed where the Southwest Indian Ocean Ridge intersects two arms of the Mid–Indian Ocean Ridge (this is the triple

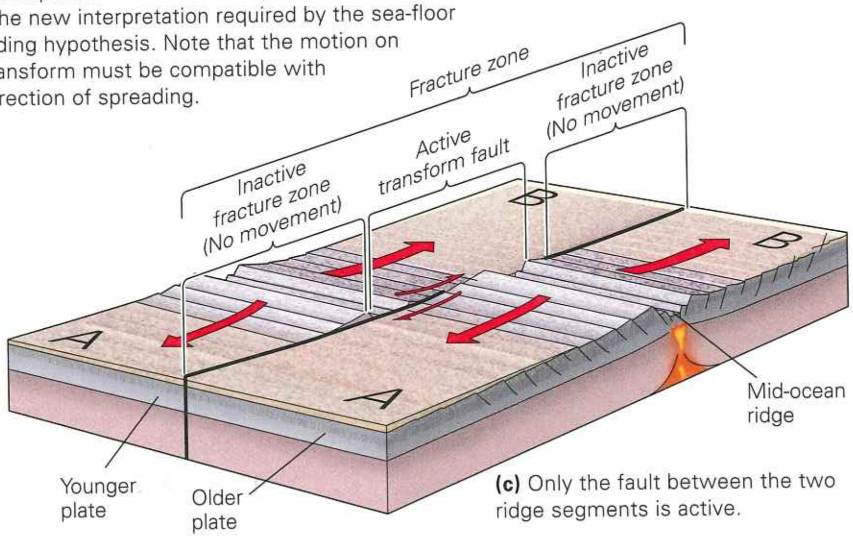
FIGURE 2.22 The concept of transform faulting.



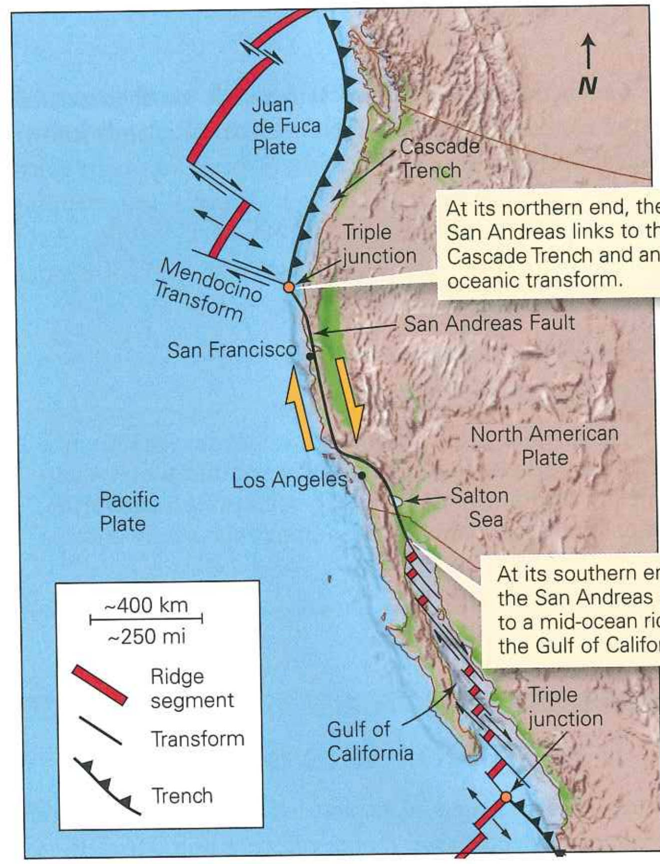
(a) Transform faults segment the Mid-Atlantic Ridge.



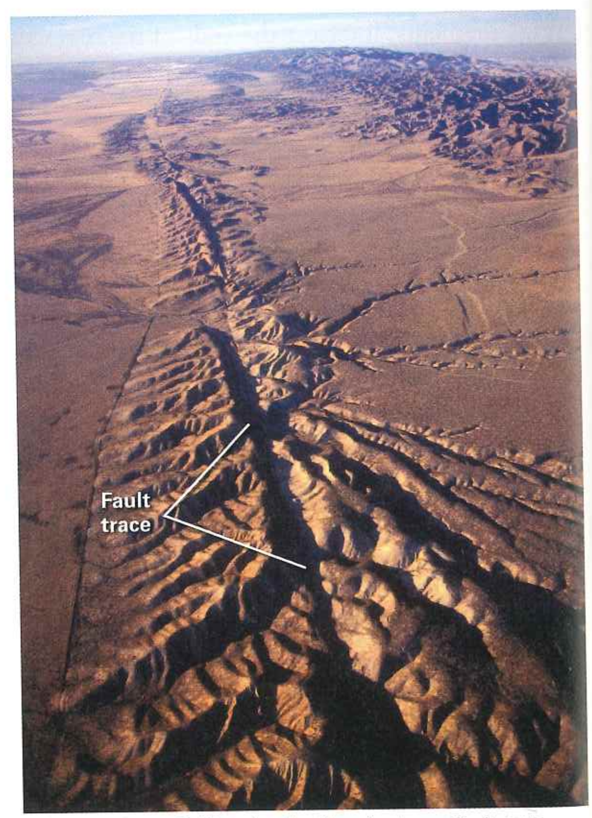
(b) A comparison of the old interpretation of transform faults with the new interpretation required by the sea-floor spreading hypothesis. Note that the motion on the transform must be compatible with the direction of spreading.



(c) Only the fault between the two ridge segments is active.

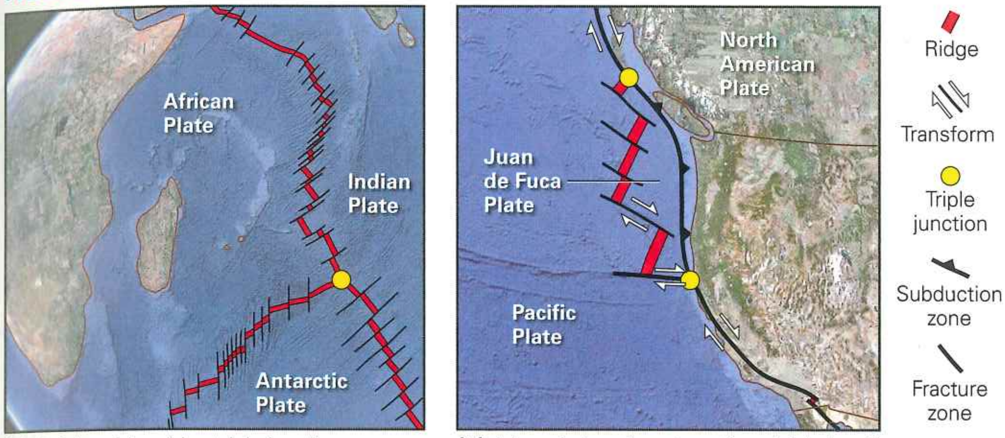


(d) The San Andreas Fault is a transform plate boundary between the North American and Pacific Plates. The Pacific is moving northwest, relative to North America.



(e) In southern California, the San Andreas Fault cuts a dry landscape. The fault trace is in the narrow valley. The land has been pushed up slightly along the fault.

FIGURE 2.23 Examples of triple junctions. The triple junctions are marked by dots.



(a) A ridge-ridge-ridge triple junction occurs in the Indian Ocean.

(b) A trench-transform-transform triple junction occurs at the north end of the San Andreas Fault.

junction of the African, Antarctic, and Australian Plates) is a ridge-ridge-ridge triple junction (**Fig. 2.23a**). The triple junction north of San Francisco is a trench-transform-transform triple junction (**Fig. 2.23b**).

Hot Spots

Most subaerial (above sea level) volcanoes are situated in the volcanic arcs that border trenches. Volcanoes also lie along mid-ocean ridges, but ocean water hides most of them. The volcanoes of volcanic arcs and mid-ocean ridges *are plate-boundary volcanoes*, in that they formed as a consequence of movement along the boundary. Not all volcanoes on Earth are plate-boundary volcanoes, however. Worldwide, geoscientists have identified about 100 volcanoes that exist as isolated points and are not a consequence of movement at a plate boundary. These are called hot-spot volcanoes, or simply **hot spots** (**Fig. 2.24**). Most hot spots are located in the interiors of plates, away from the boundaries, but a few lie along mid-ocean ridges.

What causes hot-spot volcanoes? In the early 1960s, J. Tuzo Wilson noted that *active* hot-spot volcanoes (examples that are erupting or may erupt in the future) occur at the end of a chain of *dead* volcanic islands and seamounts (formerly active volcanoes that will never erupt again). This configuration is different from that of volcanic arcs along convergent plate boundaries—at volcanic arcs, all of the volcanoes are active. With this image in mind, Wilson suggested that the position of the heat source causing a hot-spot volcano is fixed, relative to the moving plate. In Wilson's model, the active volcano represents the present-day location of the heat source, whereas the chain of dead volcanic islands represents locations on the plate that were once over the heat source but progressively moved off.

A few years later, researchers suggested that the heat source for hot spots is a **mantle plume**, a column of very hot rock rising up through the mantle to the base of the lithosphere (**Fig. 2.25a–d**). In this model, plumes originate deep in the mantle. Rock in the plume, though solid, is soft

FIGURE 2.24 The dots represent the locations of selected hot-spot volcanoes. The red lines represent hot-spot tracks. The most recent volcano (dot) is at one end of this track. Some of these volcanoes are extinct, indicating that the mantle plume no longer exists. Some hot spots are fairly recent and do not have tracks. Dashed tracks were broken by sea-floor spreading.

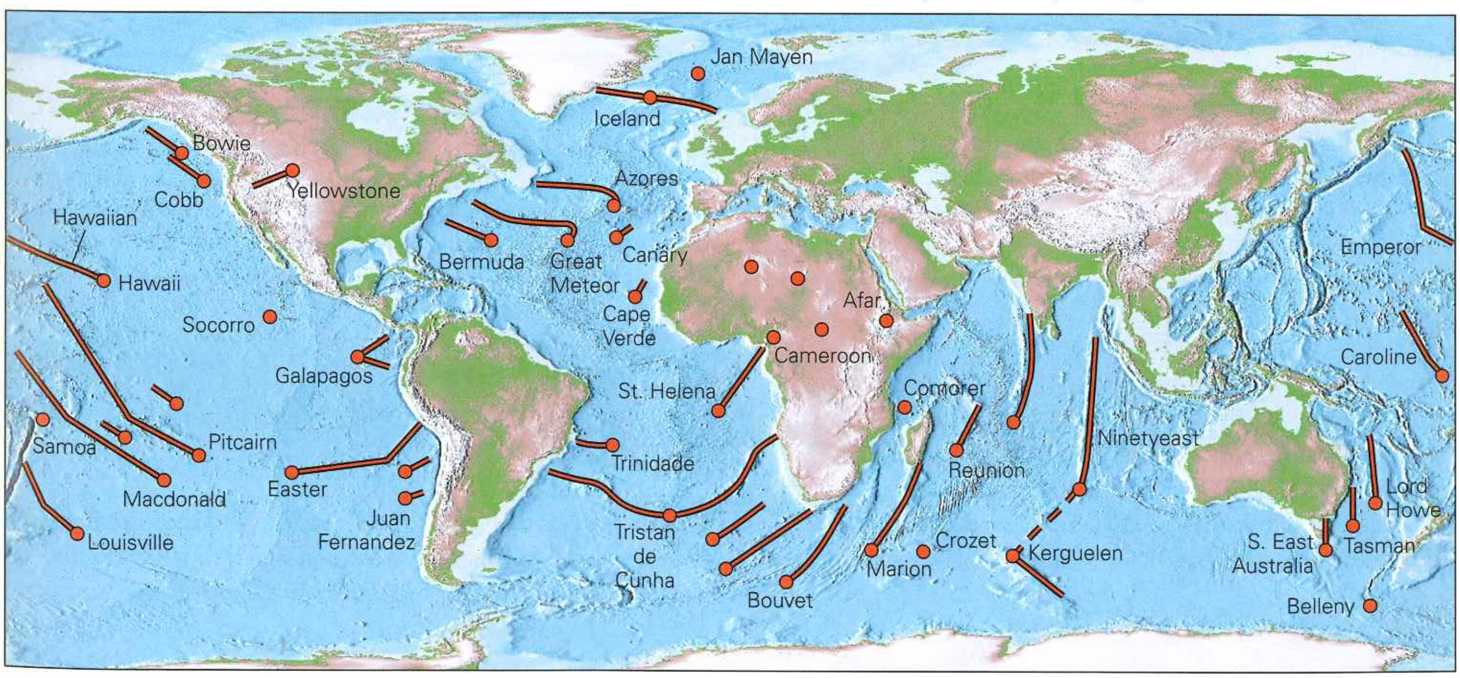
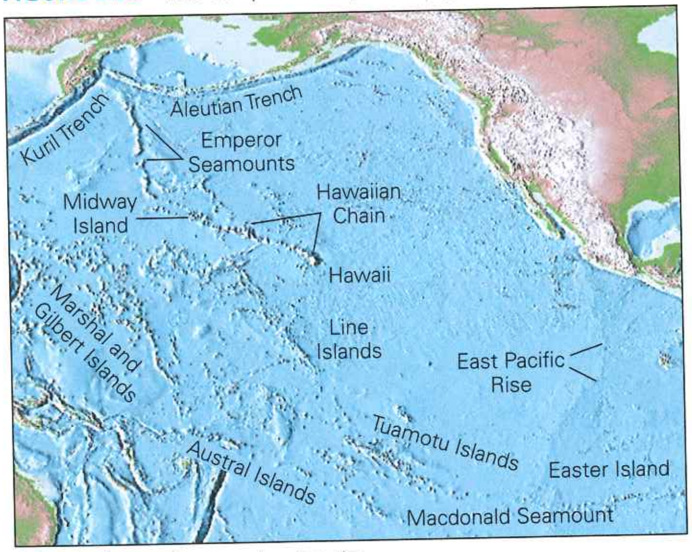
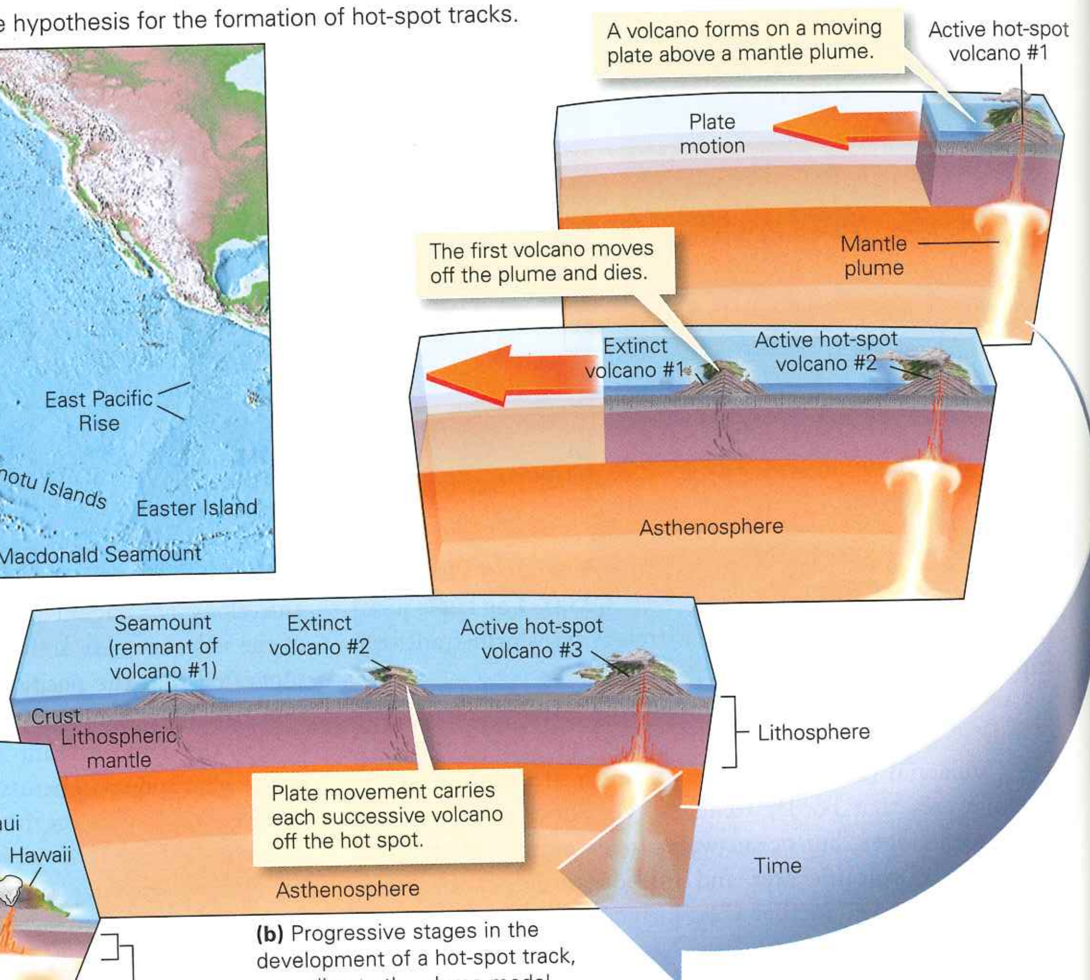


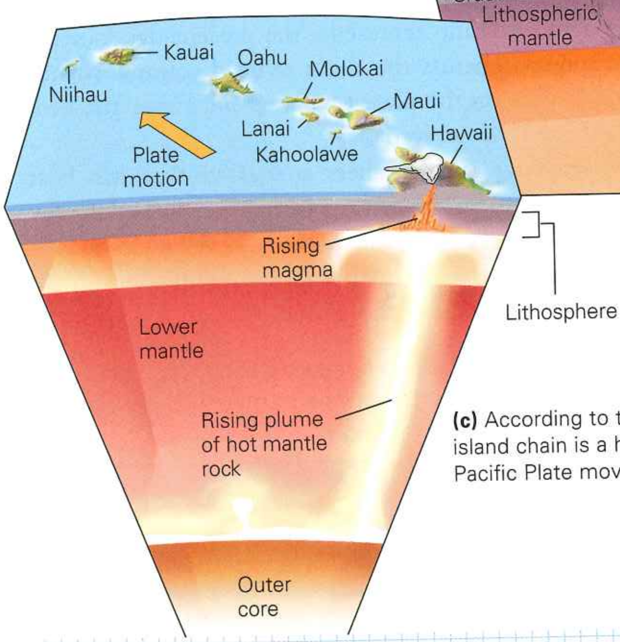
FIGURE 2.25 The deep mantle plume hypothesis for the formation of hot-spot tracks.



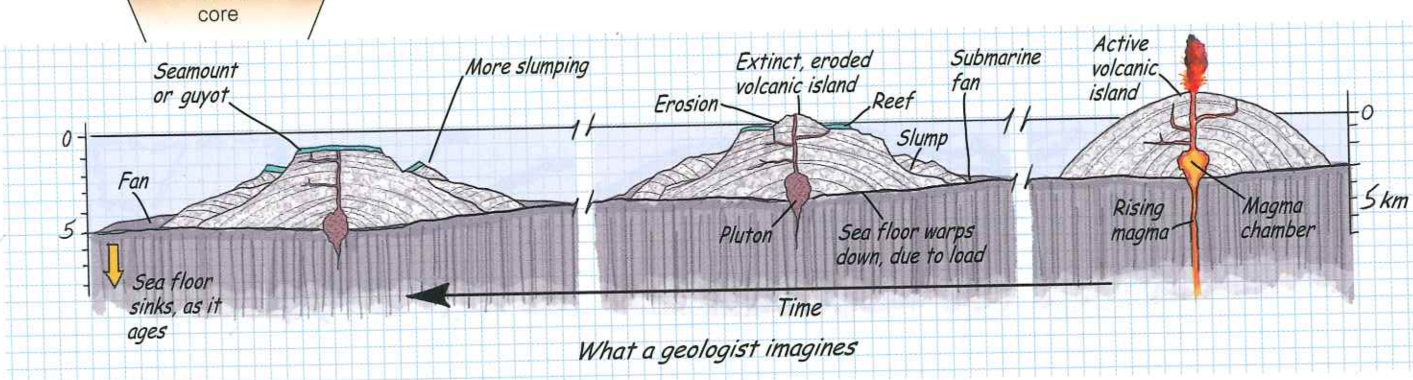
(a) A bathymetric map showing the hot-spot tracks of the Pacific Ocean.



(b) Progressive stages in the development of a hot-spot track, according to the plume model.



(c) According to the plume model, the Hawaiian island chain is a hot-spot track that formed as the Pacific Plate moved northwest relative to a plume.



(d) As a volcano moves off the hot spot, it gradually sinks below sea level, due to sinking of the plate, erosion, and slumping.

enough to flow, and rises buoyantly because it is less dense than surrounding cooler rock. When the hot rock of the plume reaches the base of the lithosphere, it partially melts (for reasons discussed in Chapter 4) and produces magma that seeps up through the lithosphere to the Earth's surface. The chain of extinct volcanoes, or **hot-spot track**, forms when the overlying plate moves over a fixed plume. This movement slowly carries the volcano off the top of the plume, so that it becomes extinct. A new, younger volcano grows over the plume.

The Hawaiian chain provides an example of the volcanism associated with a hot-spot track. Volcanic eruptions occur today only on the big island of Hawaii. Other islands to the northwest are remnants of dead volcanoes, the oldest of which is Kauai. To the northwest of Kauai, still older volcanic remnants are found. About 1,750 km northwest of Midway Island, the track bends in a more northerly direction, and the volcanic remnants no longer poke above sea level; we refer to this northerly trending segment as the Emperor seamount chain. Geologists suggest that the bend is due to a change in the direction of Pacific Plate motion at about 40 Ma.

Some hot spots lie within continents. For example, several have been active in the interior of Africa, and one now underlies Yellowstone National Park. The famous geysers (natural steam and hot-water fountains) of Yellowstone exist because hot magma, formed above the Yellowstone hot spot, lies not far below the surface of the park. While most hot spots, such as Hawaii and Yellowstone, occur in the interior of plates, away from plate boundaries, a few are positioned at points on mid-ocean ridges. The additional magma production associated with such hot spots causes a portion of the ridge to grow into a mound that can rise significantly above normal ridge-axis depths and protrude above the sea surface. Iceland, for example, is the product of hot-spot volcanism on the axis of the Mid-Atlantic Ridge.

Take-Home Message

A triple junction marks the point where three plate boundaries join. A hot spot is a place where volcanism may be due to melting at the top of a mantle plume. As a plate moves over a plume, a hot-spot track develops.

2.11 How Do Plate Boundaries Form and Die?

The configuration of plates and plate boundaries visible on our planet today has not existed for all of geologic history, and will not exist indefinitely into the future. Because of plate motion, oceanic plates form and are later consumed, while continents merge and later split apart. How does a new divergent boundary come into existence, and how does an existing convergent boundary eventually cease to exist? Most new divergent boundaries form when a continent splits and separates into two continents. We call this process **rifting**. A convergent boundary ceases to exist when a piece of buoyant lithosphere, such as a continent or an island arc, moves into the subduction zone and, in effect, jams up the system. We call this process **collision**.

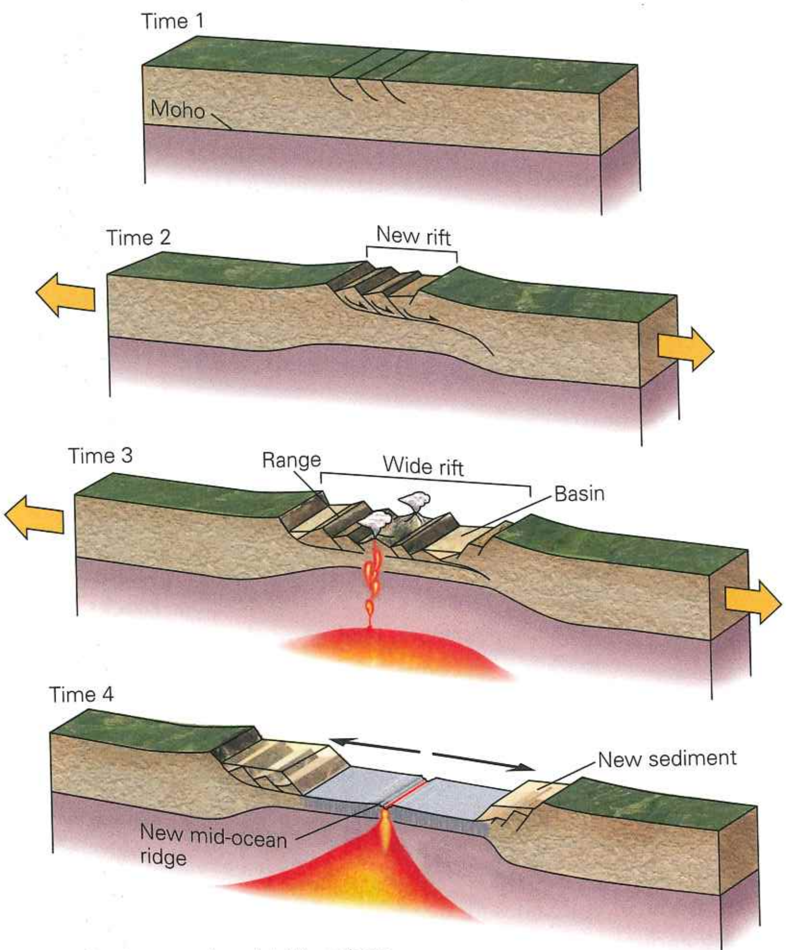
Continental Rifting

A **continental rift** is a linear belt in which continental lithosphere pulls apart (**Fig. 2.26a**). During the process, the lithosphere stretches horizontally and thins vertically, much like a piece of taffy you pull between your fingers. Nearer the surface of the continent, where the crust is cold and brittle, stretching causes rock to break and faults to develop. Blocks of rock slip down the fault surfaces, leading to the formation of a low area that gradually becomes buried by sediment. Deeper in the crust, and in the underlying lithospheric mantle, rock is warmer and softer, so stretching takes place in a plastic manner without breaking the rock. The whole region that stretches is the rift, and the process of stretching is called rifting.

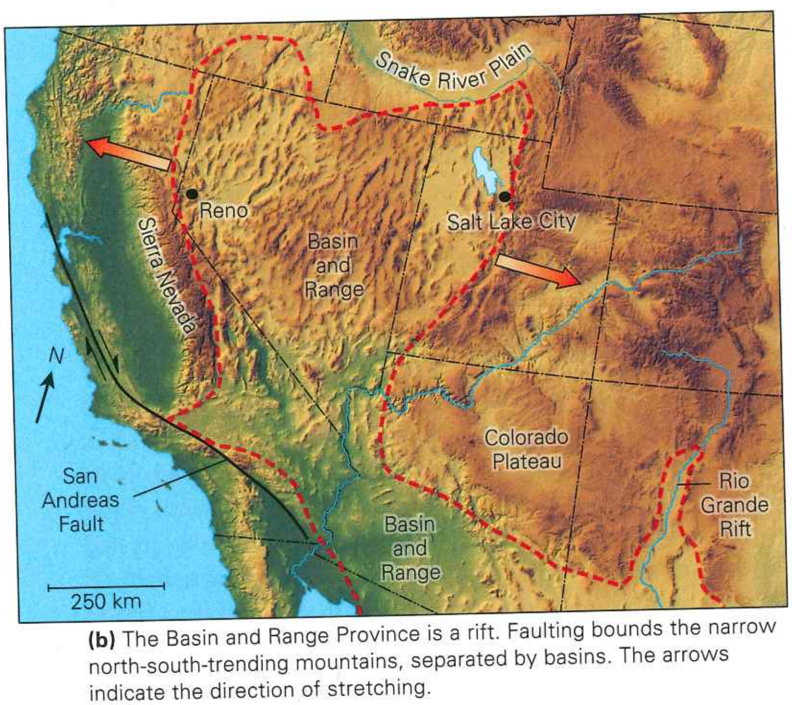
As continental lithosphere thins, hot asthenosphere rises beneath the rift and starts to melt. Eruption of the molten rock produces volcanoes along the rift. If rifting continues for a long enough time, the continent breaks in two, a new mid-ocean ridge forms, and sea-floor spreading begins. The relict of the rift evolves into a passive margin (see Fig. 2.14b). In some cases, however, rifting stops before the continent splits in two; it becomes a low-lying trough that fills with sediment. Then, the rift remains as a permanent scar in the crust, defined by a belt of faults, volcanic rocks, and a thick layer of sediment.

A major rift, known as the Basin and Range Province, breaks up the landscape of the western United States (**Fig. 2.26b**). Here, movement on numerous faults tilted blocks of crust to form narrow mountain ranges, while sediment that eroded from the blocks filled the adjacent basins (the low areas between the ranges). Another active rift slices through eastern Africa; geoscientists aptly refer to it as the East African Rift (**Fig. 2.26c, d**). To astronauts in orbit, the rift looks like a giant gash in the crust. On the ground, it consists of a deep trough bordered on both sides by high cliffs formed by faulting. Along the length of the rift, several major volcanoes smoke and fume; these include the snow-crested Mt. Kilimanjaro, towering over 6 km above the savannah. At its north end, the rift joins the Red Sea Ridge and the Gulf of Aden Ridge at a triple junction.

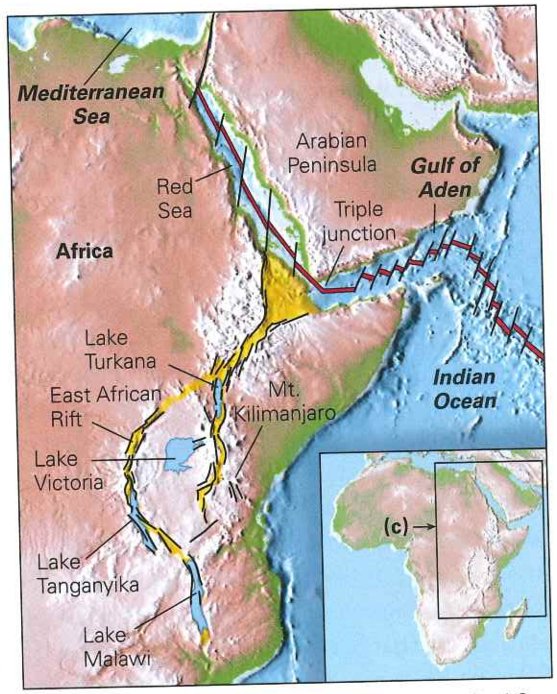
FIGURE 2.26 During the process of rifting, lithosphere stretches.



(a) When continental lithosphere stretches and thins, faulting takes place, and volcanoes erupt. Eventually, the continent splits in two and a new ocean basin forms.



(b) The Basin and Range Province is a rift. Faulting bounds the narrow north-south-trending mountains, separated by basins. The arrows indicate the direction of stretching.



(c) The East African Rift is growing today. The Red Sea started as a rift. The inset shows map locations.

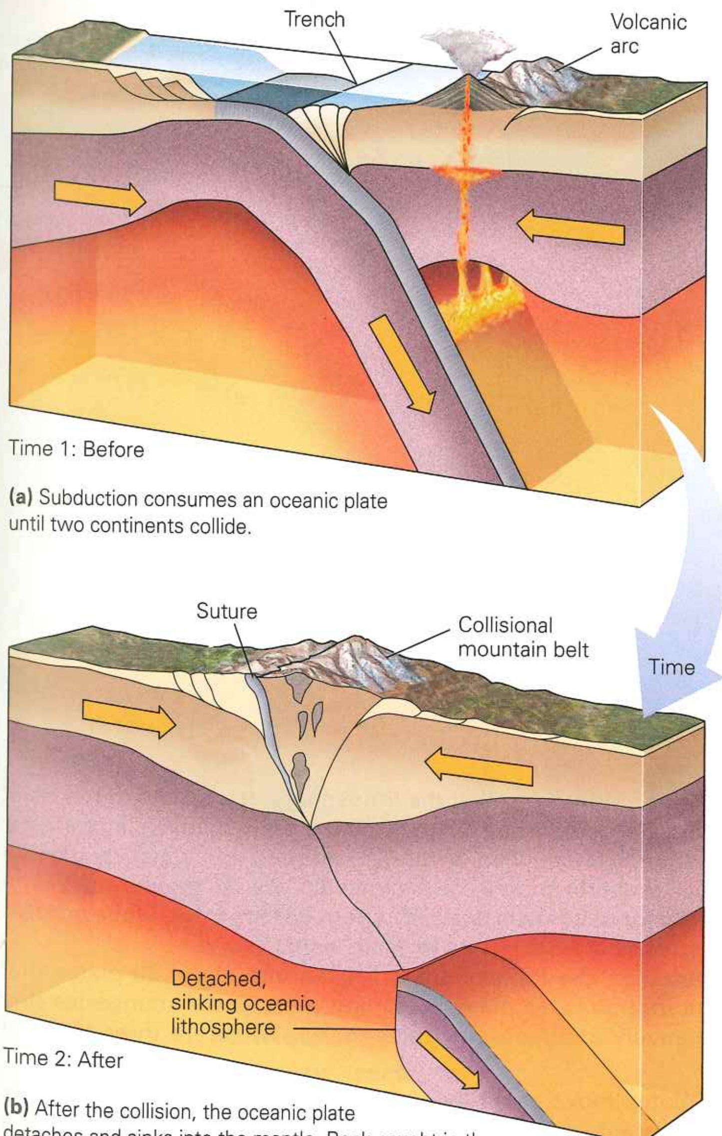
(d) An air photo of the northern end of the East African rift, showing faults and volcanoes.

Collision

India was once a small, separate continent that lay far to the south of Asia. But subduction consumed the ocean between India and Asia, and India moved northward, finally slamming into the southern margin of Asia about 40 to 50 million years ago. Continental crust, unlike oceanic crust, is too buoyant to subduct. So when India collided with Asia, the attached oceanic plate broke off and sank down into the deep mantle while India pushed hard into and partly under Asia, squeezing the rocks and sediment that once lay between the two continents into the 8-km-high welt that we now know as the Himalayan Mountains. During this process, not only did the surface of the Earth rise, but the crust became thicker. The crust beneath a collisional mountain range can be up to 60 to 70 km thick, about twice the thickness of normal continental crust. The boundary between what was once two separate continents is called a suture; slivers of ocean crust may be trapped along a suture.

Geoscientists refer to the process during which two buoyant pieces of lithosphere converge and squeeze together as collision (**Fig. 2.27a, b**). Some collisions involve two continents, whereas some involve continents and an island arc. When a collision is complete, the convergent plate boundary that once existed between the two colliding pieces ceases to exist. Collisions yield some of the most spectacular mountains on the planet, such as the Himalayas and the Alps. They also yielded major mountain ranges in the past, which subsequently eroded away so that today we see only their relicts. For example, the Appalachian Mountains in the eastern United States formed as a consequence of three collisions. After the last one, a collision between Africa and North America around 300 Ma, North America became part of the Pangaea supercontinent.

FIGURE 2.27 Continental collision (not to scale).



(a) Subduction consumes an oceanic plate until two continents collide.

(b) After the collision, the oceanic plate detaches and sinks into the mantle. Rock caught in the collision zone gets broken, bent, and squashed and forms a mountain range.

Take-Home Message

Rifting can split a continent in two and can lead to the formation of a new divergent plate boundary. When two buoyant crustal blocks, such as continents and island arcs, collide, a mountain belt forms and subduction eventually ceases.

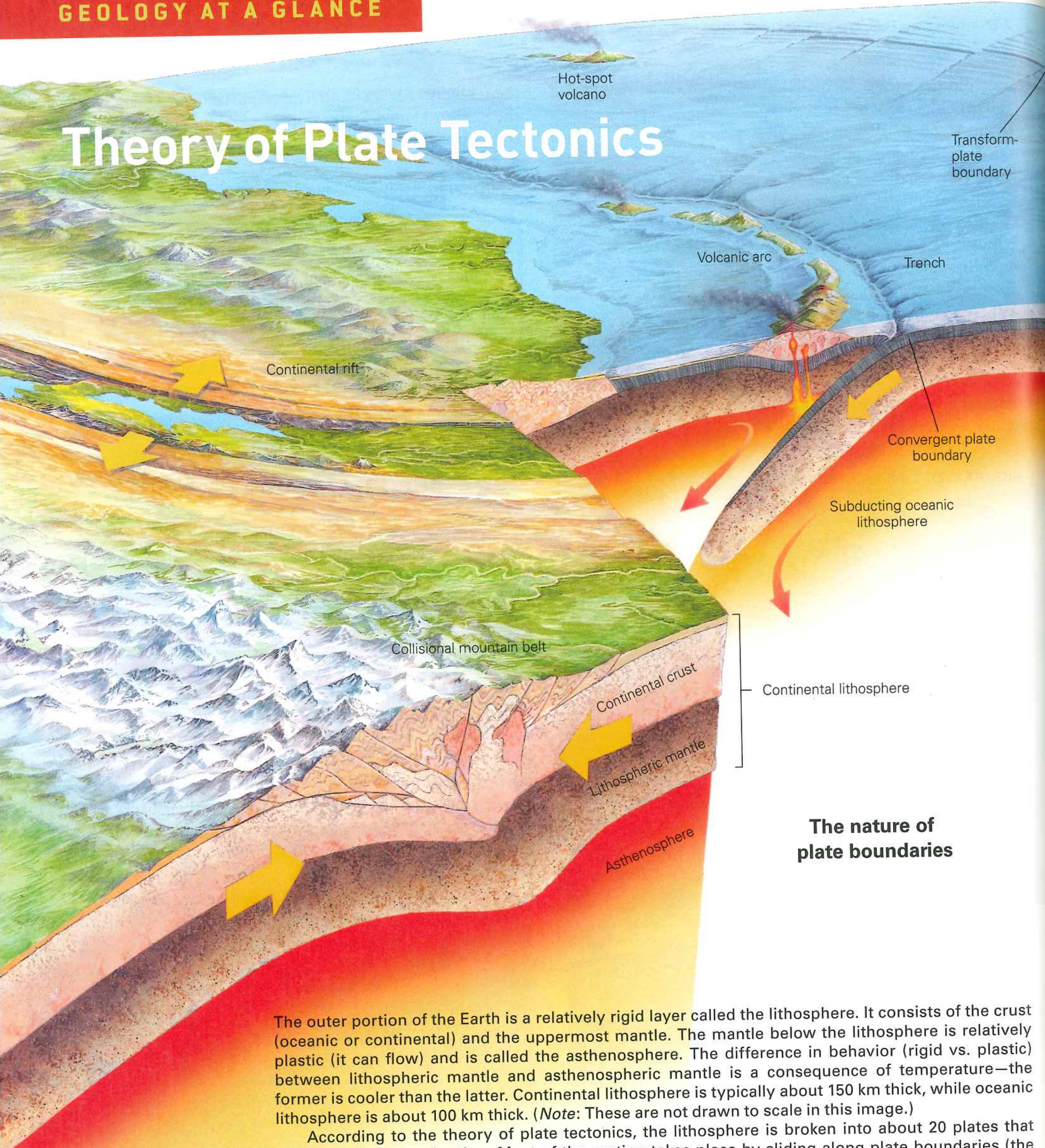
2.12 What Drives Plate Motion, and How Fast Do Plates Move?

Forces Acting on Plates

We've now discussed the many facets of plate tectonics theory (see **Geology at a Glance**, pp. 64–65). But to complete the story, we need to address a major question: "What drives plate motion?" When geoscientists first proposed plate tectonics, they thought the process occurred simply because convective flow in the asthenosphere actively dragged plates along, as if the plates were simply rafts on a flowing river. Thus, early images depicting plate motion showed simple convection cells—elliptical flow paths—in the asthenosphere. At first glance, this hypothesis looked pretty good. But, on closer examination it became clear that a model of simple convection cells carrying plates on their backs can't explain the complex geometry of plate boundaries and the great variety of plate motions that we observe on the Earth. Researchers now prefer a model in which convection, ridge push, and slab pull all contribute to driving plates. Let's look at each of these phenomena in turn.

Convection is involved in plate motions in two ways. Recall that, at a mid-ocean ridge, hot asthenosphere rises and then cools to form oceanic lithosphere which slowly moves away from the ridge until, eventually, it sinks back into the mantle at a trench. Since the material forming the plate starts out hot, cools, and then sinks, we can view the plate itself as the top of a convection cell and plate motion as a form of convection. But in this view, convection is effectively a consequence of plate motion, not the cause. Can convection actually cause plates to move? The answer may come from studies which demonstrate that the interior of the mantle, beneath the plates, is indeed convecting on a very broad scale (see Interlude D). Specifically, geologists have found that there are places where deeper, hotter asthenosphere is rising or *upwelling*, and places where shallower, colder asthenosphere is sinking or *downwelling*. Such

Theory of Plate Tectonics

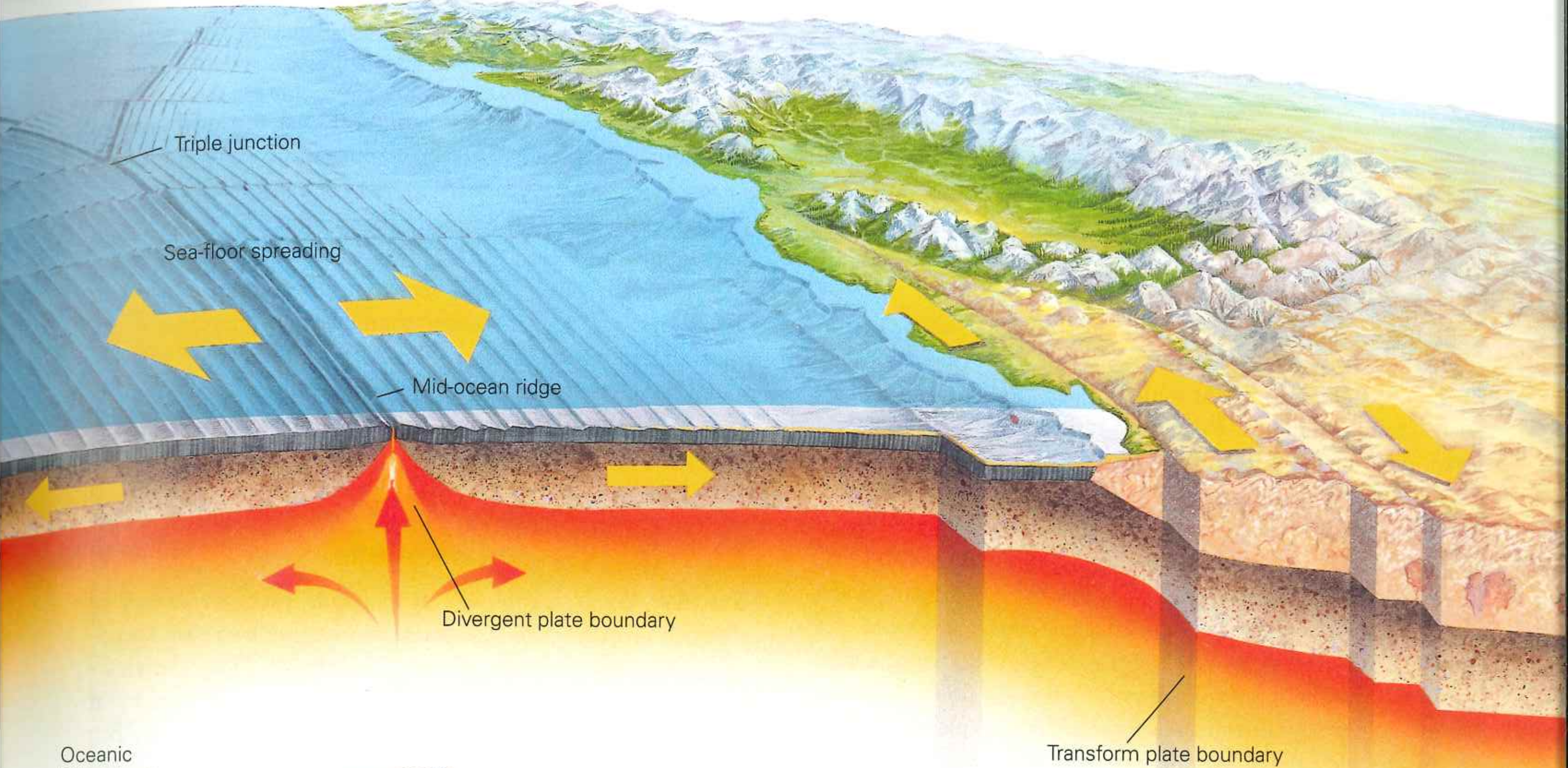


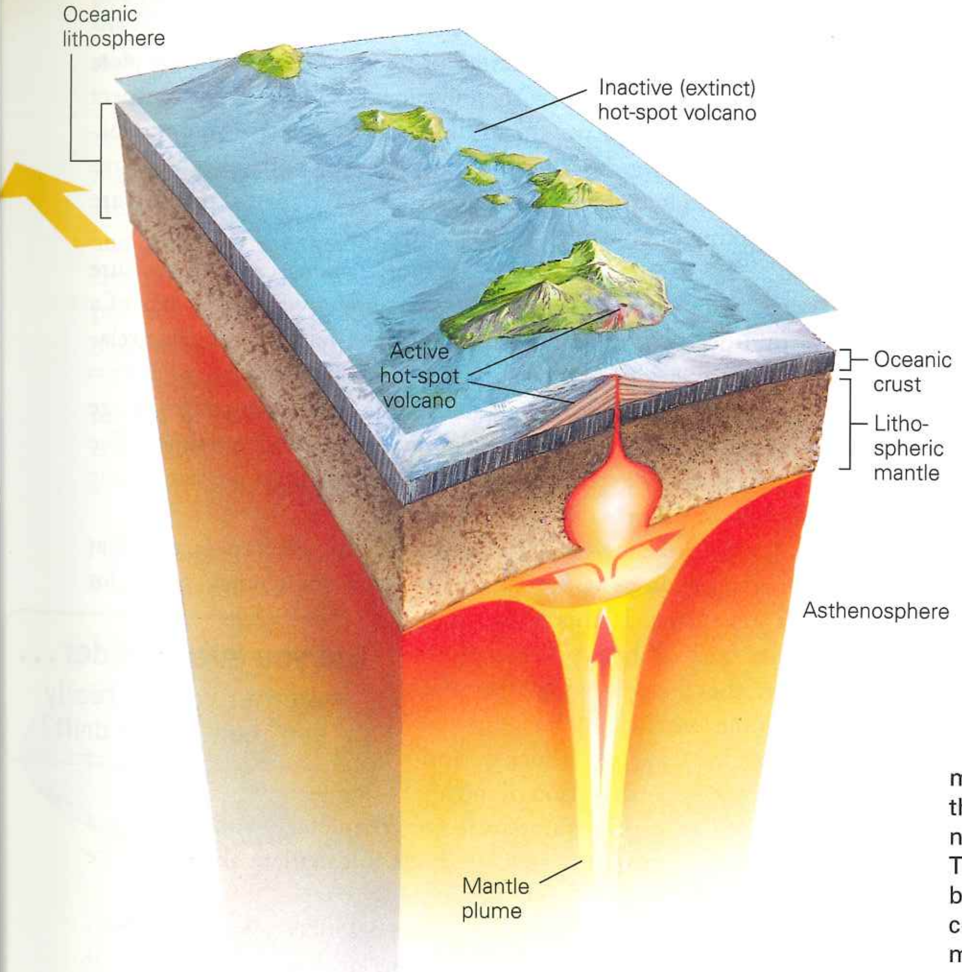
The nature of plate boundaries

The outer portion of the Earth is a relatively rigid layer called the lithosphere. It consists of the crust (oceanic or continental) and the uppermost mantle. The mantle below the lithosphere is relatively plastic (it can flow) and is called the asthenosphere. The difference in behavior (rigid vs. plastic) between lithospheric mantle and asthenospheric mantle is a consequence of temperature—the former is cooler than the latter. Continental lithosphere is typically about 150 km thick, while oceanic lithosphere is about 100 km thick. (*Note*: These are not drawn to scale in this image.)

According to the theory of plate tectonics, the lithosphere is broken into about 20 plates that move relative to each other. Most of the motion takes place by sliding along plate boundaries (the edges of plates); plate interiors stay relatively unaffected by this motion. There are three kinds of plate boundaries.

1. Divergent boundaries: Here, two plates move apart by a process called sea-floor spreading. A mid-ocean ridge delineates a divergent boundary. Asthenospheric mantle rises beneath a





Formation of a hot-spot track

mid-ocean ridge and partially melts, forming magma. The magma rises to create new oceanic crust. The lithospheric mantle thickens progressively away from the ridge axis as the plate cools.

2. Convergent boundaries: Here, two plates move together, and one plate subducts beneath another (it sinks down into the mantle). Only oceanic lithosphere can subduct. At the Earth's surface, the boundary between the two plates is marked by a deep-ocean trench. During subduction, melting above the downgoing plate produces magma that rises to form a volcanic arc.
3. Transform boundaries: Here, one plate slides sideways past another, without the creation of a new plate or the subduction of an old one. The boundary is marked by a large fault, a fracture on which sliding occurs. Transform boundaries link segments of mid-ocean ridges. They may also cut through continental lithosphere.

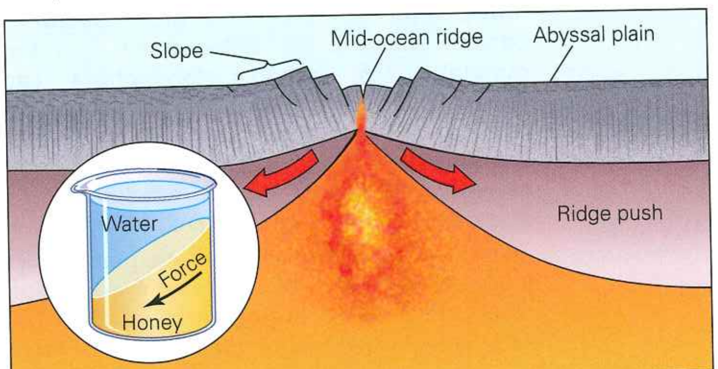
At a triple junction, three plate boundaries meet. This figure shows a triple junction where three mid-ocean ridges meet. Where two continents collide, a collisional mountain belt forms. This happens because continental crust is too buoyant to be subducted. At a continental rift, a continent stretches and may break in two. Rifts are marked by the existence of many faults. If a continent breaks apart, a new mid-ocean ridge develops.

Hot-spot volcanoes may form above plumes of hot mantle rock that rise from near the core-mantle boundary. As a plate drifts over a hot spot, it leaves a chain of extinct volcanoes.

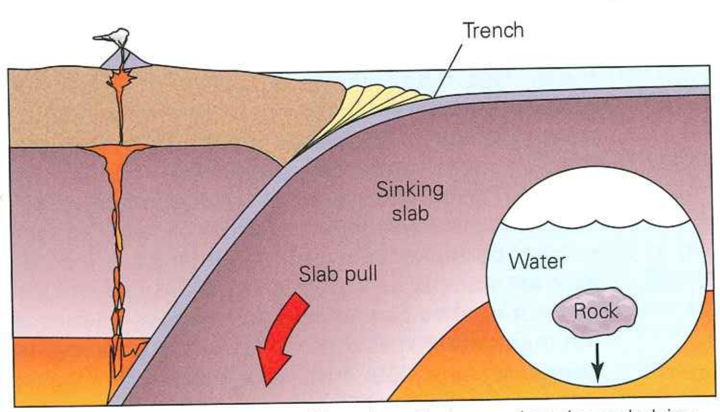
asthenospheric flow probably does exert a force on the base of plates. But the pattern of upwelling and downwelling on a global scale does not match the pattern of plate boundaries exactly. So, conceivably, asthenosphere-flow may either speed up or slow down plates depending on the orientation of the flow direction relative to the movement direction of the overlying plate.

Ridge-push force develops simply because the lithosphere of mid-ocean ridges lies at a higher elevation than that of the adjacent abyssal plains (**Fig. 2.28a**). To understand ridge-push force, imagine you have a glass containing a layer of water over a layer of honey. By tilting the glass momentarily and then returning it to its upright position, you can create a temporary slope in the boundary between these substances. While the boundary has this slope, gravity causes the weight of elevated honey to push against the glass adjacent to the side where the honey surface lies at lower elevation. The geometry of a mid-ocean ridge resembles this situation, for sea floor of a mid-ocean ridge is higher than sea floor of abyssal plains. Gravity causes the elevated lithosphere at the ridge axis to push on the lithosphere that lies farther from the axis, making it move away. As lithosphere moves away from the ridge axis, new hot asthenosphere rises to fill the gap. Note that the local upward movement of asthenosphere beneath a mid-ocean ridge is a *consequence* of sea-floor spreading, not the cause.

FIGURE 2.28 Forces driving plate motions. Both ridge push and slab pull make plates move.



(a) Ridge push develops because the region of a rift is elevated. Like a wedge of honey with a sloping surface, the mass of the ridge pushes sideways.



(b) Slab pull develops because lithosphere is denser than the underlying asthenosphere, and sinks like a stone in water (though much more slowly).

Slab-pull force, the force that subducting, downgoing plates apply to oceanic lithosphere at a convergent margin, arises simply because lithosphere that was formed more than 10 million years ago is denser than asthenosphere, so it can sink into the asthenosphere (**Fig. 2.28b**). Thus, once an oceanic plate starts to sink, it gradually pulls the rest of the plate along behind it, like an anchor pulling down the anchor line. This "pull" is the slab-pull force.

The Velocity of Plate Motions

How fast do plates move? It depends on your frame of reference. To illustrate this concept, imagine two cars speeding in the same direction down the highway. From the viewpoint of a tree along the side of the road, Car A zips by at 100 km an hour, while Car B moves at 80 km an hour. But relative to Car B, Car A moves at only 20 km an hour. Geologists use two different frames of reference for describing plate velocity. If we describe the movement of Plate A with respect to Plate B, then we are speaking about **relative plate velocity**. But if we describe the movement of both plates relative to a fixed location in the mantle below the plates, then we are speaking of **absolute plate velocity** (**Fig. 2.29**).

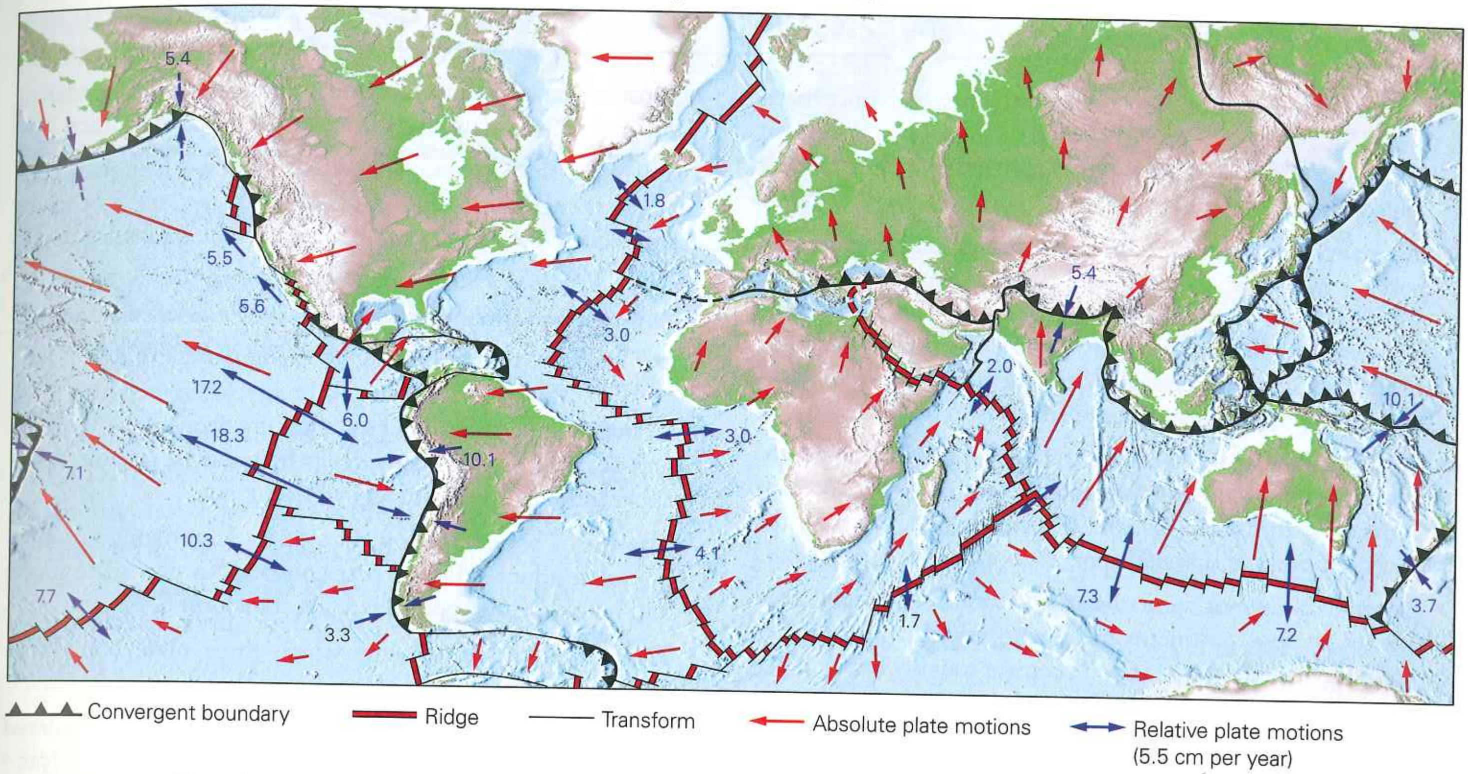
To determine relative plate motions, geoscientists measure the distance of a known magnetic anomaly from the axis of a mid-ocean ridge and then calculate the velocity of a plate relative to the ridge axis by applying this equation: plate velocity = distance from the anomaly to the ridge axis divided by the age of the anomaly (velocity, by definition, is distance ÷ time). The velocity of the plate on one side of the ridge relative to the plate on the other is twice this value.

To estimate absolute plate motions, we can *assume* that the position of a mantle plume does not change much for a long time. If this is so, then the track of hot-spot volcanoes on the plate moving over the plume provides a record of the plate's absolute velocity and indicates the direction of movement. (In reality, plumes are not completely fixed; geologists use other, more complex methods to calculate absolute plate motions.)

Did you ever wonder... whether we can really "see" continents drift?

Working from the calculations described above, geologists have determined that plate motions on Earth today occur at rates of about 1 to 15 cm per year—about the rate that your fingernails grow. But these rates, though small, can yield large displacements given the immensity of geologic time. At a rate of 10 cm per year, a plate can move 100 km in a million years! Can we detect such slow rates? Until the last decade, the

FIGURE 2.29 *Relative plate velocities*: The blue arrows show the rate and direction at which the plate on one side of the boundary is moving with respect to the plate on the other side. The length of an arrow represents the velocity. *Absolute plate velocities*: The red arrows show the velocity of the plates with respect to a fixed point in the mantle.



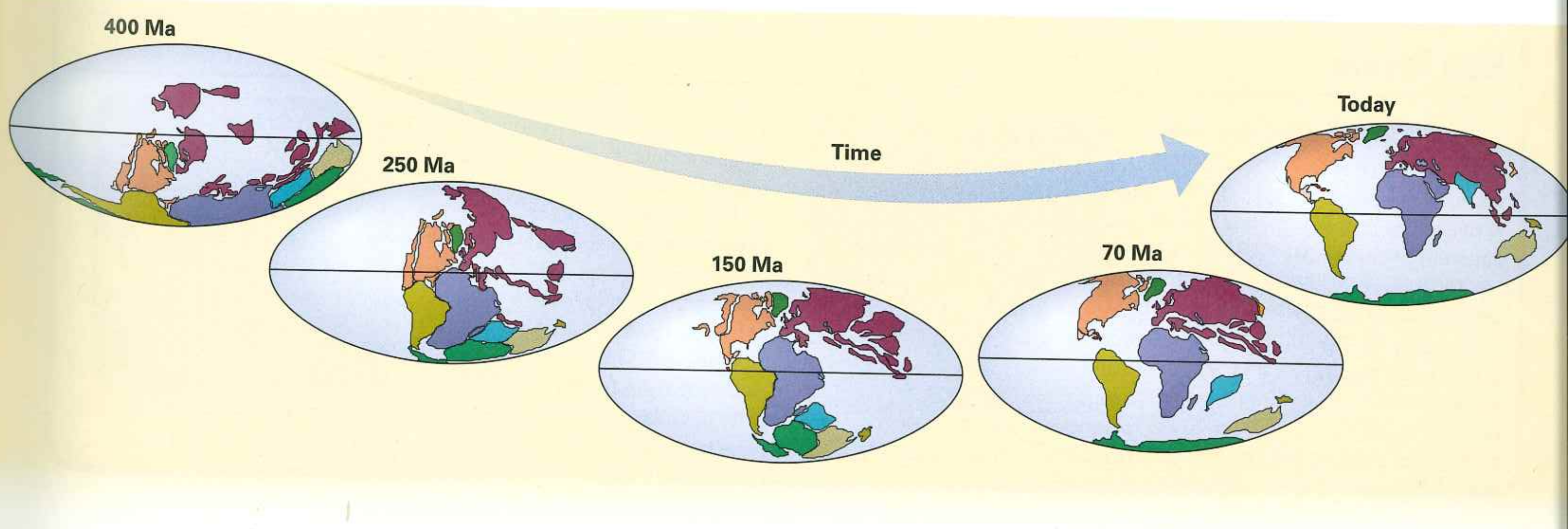
answer was no. Now the answer is yes, because of satellites orbiting the Earth with **global positioning system (GPS)** technology. Automobile drivers use GPS receivers to find their destinations, and geologists use them to monitor plate motions. If we calculate carefully enough, we can detect displacements of millimeters per year. In other words, we can now see the plates move—this observation serves as the ultimate proof of plate tectonics.

Taking into account many data sources that define the motion of plates, geologists have greatly refined the image of continental drift that Wegener tried so hard to prove nearly a century ago. We can now see how the map of our planet's surface has evolved radically during the past 400 million years (**Fig. 2.30**), and even before.

Take-Home Message

Plates move at 1 to 15 cm/yr. Relative motion specifies the rate that a plate moves relative to its neighbor, whereas absolute motion specifies the rate that a plate moves relative to a fixed point beneath the plate. GPS measurements can detect relative plate motions directly.

FIGURE 2.30 Due to plate tectonics, the map of Earth's surface slowly changes. Here we see the assembly, and later the breakup, of Pangaea during the past 400 million years.



Chapter Summary

- Alfred Wegener proposed that continents had once been joined together to form a single huge supercontinent (Pangaea) and had subsequently drifted apart. This idea is the continental drift hypothesis.
- Wegener drew from several different sources of data to support his hypothesis: (1) the correlation of coastlines; (2) the distribution of late Paleozoic glaciers; (3) the distribution of late Paleozoic climatic belts; (4) the distribution of fossil species; and (5) correlation of distinctive rock assemblages now on opposite sides of the ocean.
- Rocks retain a record of the Earth's magnetic field that existed at the time the rocks formed. This record is called paleomagnetism. By measuring paleomagnetism in successively older rocks, geologists discovered apparent polar-wander paths.
- Apparent polar-wander paths are different for different continents, because continents move with respect to each other, while the Earth's magnetic poles remain roughly fixed.
- Around 1960, Harry Hess proposed the hypothesis of sea-floor spreading. According to this hypothesis, new sea floor forms at mid-ocean ridges, then spreads symmetrically away from the ridge axis. Eventually, the ocean floor sinks back into the mantle at deep-ocean trenches.
- Geologists documented that the Earth's magnetic field reverses polarity every now and then. The record of reversals is called the magnetic-reversal chronology.
- A proof of sea-floor spreading came from the interpretation of marine magnetic anomalies and from drilling studies which proved that sea floor gets progressively older away from a mid-ocean ridge.
- The lithosphere is broken into discrete plates that move relative to each other. Continental drift and sea-floor spreading are manifestations of plate movement.
- Most earthquakes and volcanoes occur along plate boundaries; the interiors of plates remain relatively rigid and intact.
- There are three types of plate boundaries—divergent, convergent, and transform—distinguished from each other by the movement the plate on one side of the boundary makes relative to the plate on the other side.
- Divergent boundaries are marked by mid-ocean ridges. At divergent boundaries, sea-floor spreading produces new oceanic lithosphere.
- Convergent boundaries are marked by deep-ocean trenches and volcanic arcs. At convergent boundaries, oceanic lithosphere subducts beneath an overriding plate.
- Transform boundaries are marked by large faults at which one plate slides sideways past another. No new plate forms and no old plate is consumed at a transform boundary.
- Triple junctions are points where three plate boundaries intersect.
- Hot spots are places where volcanism occurs at an isolated volcano. As a plate moves over the hot spot, the volcano moves off and dies, and a new volcano forms over the hot spot. Hot spots may be caused by mantle plumes.
- A large continent can split into two smaller ones by the process of rifting. During rifting, continental lithosphere stretches and thins. If it finally breaks apart, a new mid-ocean ridge forms and sea-floor spreading begins.
- Convergent plate boundaries cease to exist when a buoyant piece of crust (a continent or an island arc) moves into the subduction zone. When that happens, collision occurs.
- Ridge-push force and slab-pull force contribute to driving plate motions. Plates move at rates of about 1 to 15 cm per year. Modern satellite measurements can detect these motions.

Key Terms

absolute plate velocity (p. 66)
abyssal plain (p. 43)
accretionary prism (p. 55)
active margin (p. 51)
apparent polar-wander path (p. 42)
asthenosphere (p. 50)
bathymetry (p. 43)
black smoker (p. 52)
chron (p. 47)
collision (p. 61)
continental drift (p. 35)
continental rift (p. 61)
continental shelf (p. 52)
convergent boundary (p. 52)
divergent boundary (p. 52)
fracture zone (pp. 43, 57)
global positioning system (GPS) (p. 67)
hot spot (p. 59)
hot-spot track (p. 61)
lithosphere (p. 50)
lithosphere plate (p. 51)
magnetic anomaly (p. 46)
magnetic declination (p. 40)
magnetic dipole (p. 40)
magnetic inclination (p. 41)
magnetic pole (p. 40)
magnetic reversal (p. 46)
mantle plume (p. 59)
marine magnetic anomaly (p. 46)
mid-ocean ridge (p. 43)
paleomagnetism (p. 41)
paleopole (p. 42)
Pangaea (p. 35)
passive margin (p. 51)
plate boundary (p. 51)
plate tectonics (p. 36)
relative plate velocity (p. 66)
ridge-push force (p. 66)
rifting (p. 61)
sea-floor spreading (pp. 36, 45)
seamount (p. 43)
slab-pull force (p. 66)
subduction (pp. 36, 54)
transform boundary (pp. 52, 57)
trench (pp. 43, 54)
triple junction (p. 57)
volcanic arc (pp. 43, 55)
Wadati-Benioff zone (p. 55)

smartwork smartwork.wwnorton.com

Every chapter of SmartWork contains active learning exercises to assist you with reading comprehension and concept mastery. This chapter also features:

- Animation exercises on plate movements, subduction, and hot spots.
- A video exercise on divergent plate boundaries.
- Problems that help students determine relative plate velocities.

Review Questions

1. What was Wegener's continental drift hypothesis? What was his evidence? Why didn't other geologists agree?
2. How do apparent polar-wander paths show that the continents, rather than the poles, have moved?
3. Describe the hypothesis of sea-floor spreading.
4. Describe the pattern of marine magnetic anomalies across a mid-ocean ridge. How is this pattern explained?
5. How did drilling into the sea floor contribute further proof of sea-floor spreading? How did the sea-floor-spreading hypothesis explain variations in ocean floor heat flow?
6. What are the characteristics of a lithosphere plate? Can a single plate include both continental and oceanic lithosphere?
7. How does oceanic lithosphere differ from continental lithosphere in thickness, composition, and density?
8. How do we identify a plate boundary?
9. Describe the three types of plate boundaries.
10. How does crust form along a mid-ocean ridge?
11. Why is the oldest oceanic lithosphere less than 200 Ma?
12. Describe the major features of a convergent boundary.
13. Why are transform plate boundaries required on an Earth with spreading and subducting plate boundaries?
14. What is a triple junction?
15. How is a hot-spot track produced, and how can hot-spot tracks be used to track the past motions of a plate?
16. Describe the characteristics of a continental rift and give examples of where this process is occurring today.
17. Describe the process of continental collision and give examples of where this process has occurred.
18. Discuss the major forces that move lithosphere plates.
19. Explain the difference between relative plate velocity and absolute plate velocity.

On Further Thought

20. Why are the marine magnetic anomalies bordering the East Pacific Rise in the Pacific Ocean wider than those bordering the Mid-Atlantic Ridge?
21. The North Atlantic Ocean is 3,600 km wide. Sea-floor spreading along the Mid-Atlantic Ridge occurs at 2 cm per year. When did rifting start to open the Atlantic?

SEE FOR YOURSELF B... Plate Tectonics

Download the *Google Earth*™ from the Web in order to visit the locations described below (instructions appear in the Preface of this book). You'll find further locations and associated active-learning exercises on Worksheet B of our **Geotours Workbook**.

The South Atlantic
Latitude 11°38′14.93″S,
Longitude 14°12′57.84″W

A view of the South Atlantic from 13,500 km emphasizes that South America's coast looks like it could fit tightly against Africa's. If you rotate the Earth, you'll see that the east coast of the United States could fit against Africa's northwest coast.

Triple Junction, Japan
Latitude 37°56′27.58″N,
Longitude 140°28′53.87″E

A space view of Japan's coast from 3,250 km shows the presence of deep-sea trenches and a broad accretionary prism. The Pacific is subducting beneath Japan. A triple junction of three trenches lies along the east coast.