
KILO MOANA

Divergent Boundaries: Origin and Evolution of the Ocean Floor

CHAPTER

13

The University of Hawaii's research vessel, Kilo Moana, *arriving at Kodiak, Alaska. (Photo by Marshalena Delaney/AP Photo/US Coast Guard)*

The ocean is the largest feature on Earth, covering more than 70 percent of our planet's surface. One of the main reasons that Wegener's continental drift hypothesis was not widely accepted when first proposed was that so little was known about the ocean floor. Until the 20th century, investigators used weighted lines to measure water depth. In deep water these depth measurements, or soundings, took hours to perform and could be wildly inaccurate.

With the development of new marine tools following World War II, our knowledge of the diverse topography of the ocean floor grew rapidly. One of the most interesting discoveries was the global oceanic ridge system. This broad elevated landform, which stands 2 to 3 kilometers above the adjacent deep-ocean basins, is the longest topographic feature on Earth.

Today we know that oceanic ridges mark divergent plate margins where new oceanic lithosphere originates. We also know that deep-ocean trenches represent convergent plate boundaries, where oceanic lithosphere is subducted into the mantle. Because the process of plate tectonics is creating oceanic crust at mid-ocean ridges and consuming it at subduction zones, the oceanic crust is continually being renewed and recycled.

In this chapter, we will examine the topography of the ocean floor and look at the processes that produced its varied features. You will also learn about the composition, structure, and origin of oceanic crust. In addition, you will examine those processes that recycle oceanic lithosphere and consider how this activity causes Earth's landmasses to migrate about the face of the globe.

An Emerging Picture of the Ocean Floor

GEODe EARTH
Divergent Boundaries
▸ Mapping the Ocean Floor

If all water were drained from the ocean basins, a great variety of features would be seen, including broad volcanic peaks, deep trenches, extensive plains, linear mountain chains, and large plateaus. In fact, the scenery would be nearly as diverse as that on the continents.

An understanding of seafloor features came with the development of techniques that measure the depth of the oceans. **Bathymetry** (*bathos* = depth, *metry* = measurement) is the measurement of ocean depths and the charting of the shape or topography of the ocean floor.

Mapping the Seafloor

The first understanding of the ocean floor's varied topography did not unfold until the historic three-and-a-half-year voyage of the HMS *Challenger* (Figure 13.1). From December 1872 to May 1876, the *Challenger* expedition made the first—and perhaps still most comprehensive—study of the global ocean ever attempted by one agency. The 127,500-kilometer (79,200-mile) trip took the ship and its crew of scientists to every ocean except the Arctic. Throughout the voyage, they sampled a multitude of ocean properties, including water depth, which was accomplished by laboriously lowering a long weighted line overboard. Not many years later, the knowledge gained by the *Challenger* of the ocean's great depth and varied topography was further expanded with the laying of transatlantic communication cables, especially in the North Atlantic Ocean. However, as long as a weighted line was the only way to measure ocean depths, knowledge of seafloor features remained extremely limited.

Bathymetric Techniques Today sound energy is used to measure water depths. The basic approach employs some type of **sonar,** an acronym for *so*und *na*vigation and *r*anging. The first devices that used sound to measure water depth, called **echo sounders,** were developed early in the twentieth century. Echo sounders work by transmitting a sound wave (called a *ping*) into the water in order to produce an echo when it bounces off any object, such as a large marine organism or the ocean floor (Figure 13.2A). A sensitive receiver intercepts the echo reflected from the bottom, and a clock precisely measures the travel time to fractions of a second. By knowing the speed of sound waves in water—about 1500 meters (4900 feet) per second—and the time required for the energy pulse to reach the ocean floor and

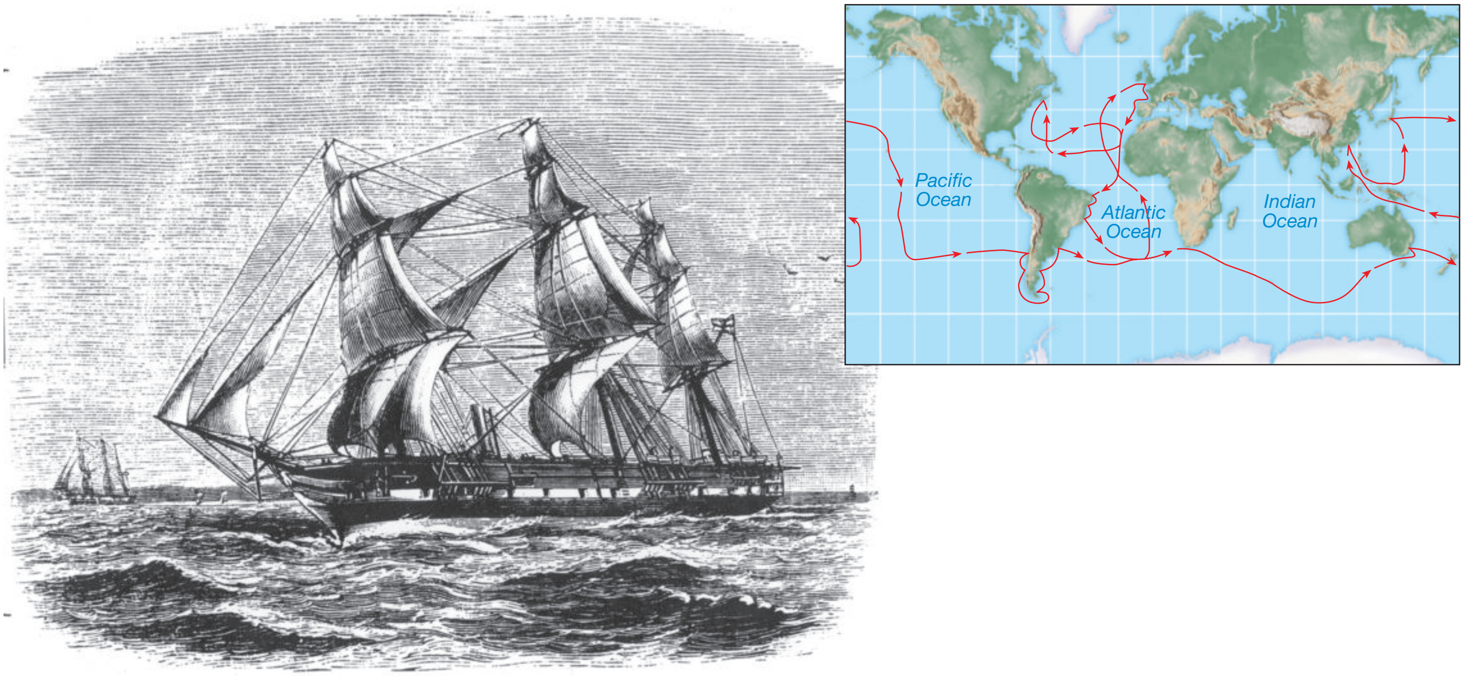


FIGURE 13.1 The first systematic bathymetric measurements of the ocean were made aboard the H.M.S. *Challenger* during its historic three-and-a-half-year voyage. Inset shows route of the H.M.S. *Challenger,* which departed England in December of 1872 and returned in May 1876. (From C.W. Thompson and Sir John Murray, *Report on the Scientific Results of the Voyage of the H.M.S. Challenger,* Vol. 1, Great Britain: Challenger Office, 1895, Plate 1. Library of Congress)

return, depth can be calculated. The depths determined from continuous monitoring of these echoes are plotted so a profile of the ocean floor is obtained. By laboriously combining profiles from several adjacent traverses, a chart of the seafloor can be produced.

Following World War II, the U.S. Navy developed *sidescan sonar* to look for mines and other explosive devices. These torpedo-shaped instruments can be towed behind a ship where they send out a fan of sound extending to either side of the ship's track. By combining swaths of sidescan

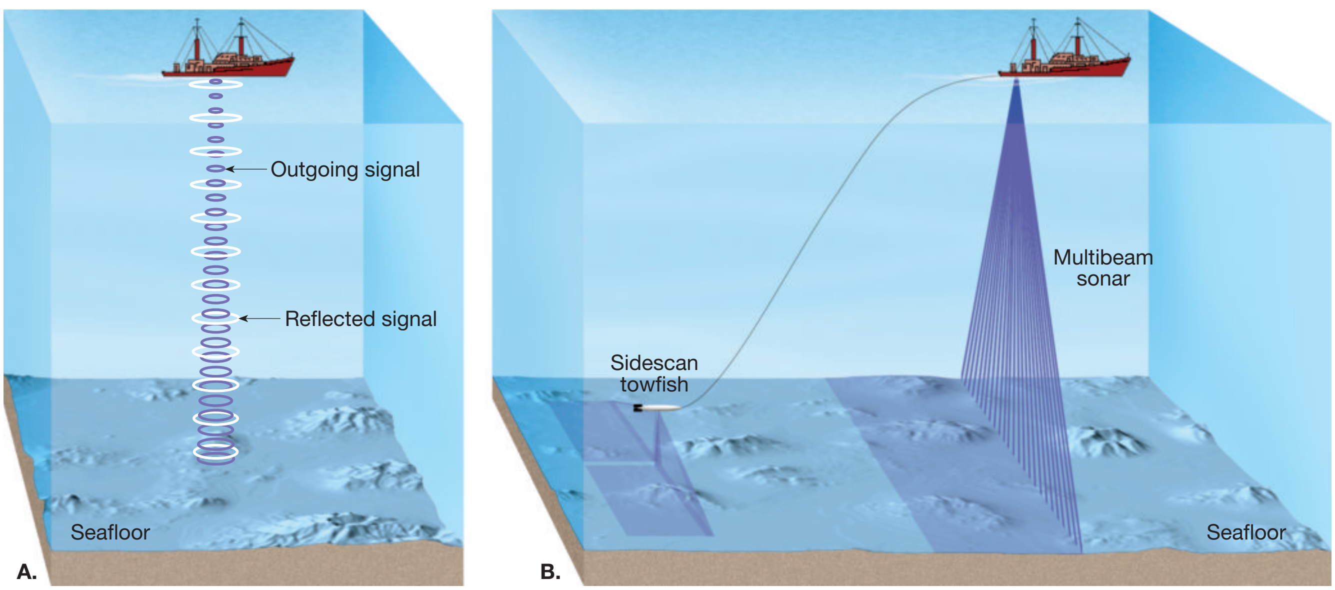


FIGURE 13.2 Various types of sonar. **A.** An echo sounder determines the water depth by measuring the time interval required for an acoustic wave to travel from a ship to the seafloor and back. The speed of sound in water is 1500 m/sec. Therefore, depth $= \frac{1}{2}$(1500 m/sec $\times$ echo travel time). **B.** Modern multibeam sonar and sidescan sonar obtain an "image" of a narrow swath of seafloor every few seconds.

sonar data, researchers produced the first photograph-like images of the seafloor. Although sidescan sonar provides valuable views of the seafloor, it does not provide bathymetric (water depth) data.

This problem is not present in the *high-resolution multibeam* instruments developed during the 1990s. These systems use hull-mounted sound sources that send out a fan of sound, then record reflections from the seafloor through a set of narrowly focused receivers aimed at different angles. Thus, rather than obtaining the depth of a single point every few seconds, this technique makes it possible for a survey ship to map the features of the ocean floor along a strip tens of kilometers wide (Figure 13.3). When a ship uses multibeam sonar to make a map of a section of seafloor, it travels through the area in a regularly spaced back-and-forth pattern known as "mowing the lawn." Furthermore, these systems can collect bathymetric data of such high resolution that they can distinguish depths that differ by less than a meter.

Despite their greater efficiency and enhanced detail, research vessels equipped with multibeam sonar travel at a mere 10 to 20 kilometers (6 to 12 miles) per hour. It would take at least 100 vessels outfitted with this equipment hundreds of years to map the entire seafloor. This explains why only about 5 percent of the seafloor has been mapped in detail—and why large areas of the seafloor have not yet been mapped with sonar at all.

Seismic Reflection Profiles Marine geologists are also interested in viewing the rock structure beneath the sediments that blanket much of the seafloor. This can be accomplished by making a **seismic reflection profile.** To construct such a profile, strong low-frequency sounds are produced by explosions (depth charges) or air guns. These sound waves penetrate beneath the seafloor and reflect off the contacts between rock layers and fault zones, just like sonar reflects off the bottom of the sea. Figure 13.4 shows a seismic profile of a portion of the Madeira abyssal plain in the eastern Atlantic. Although the seafloor is flat, notice the irregular ocean crust buried by a thick accumulation of sediments.

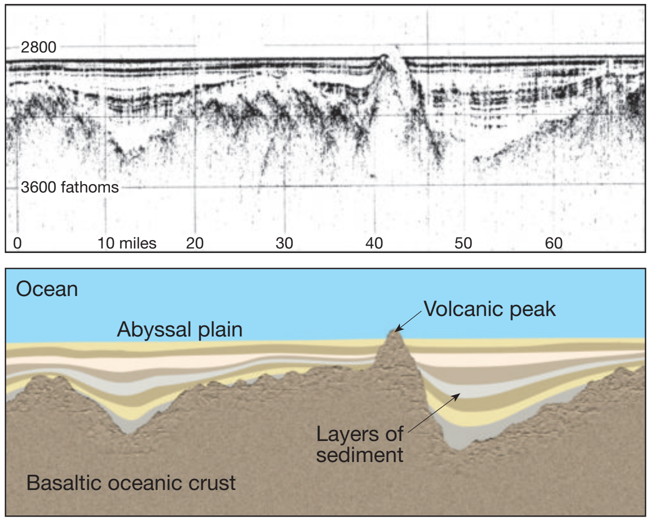


FIGURE 13.4 Seismic cross section and matching sketch across a portion of the Madeira abyssal plain in the eastern Atlantic Ocean, showing the irregular oceanic crust buried by sediments. (Image courtesy of Charles Hollister, Woods Hole Oceanographic Institution)

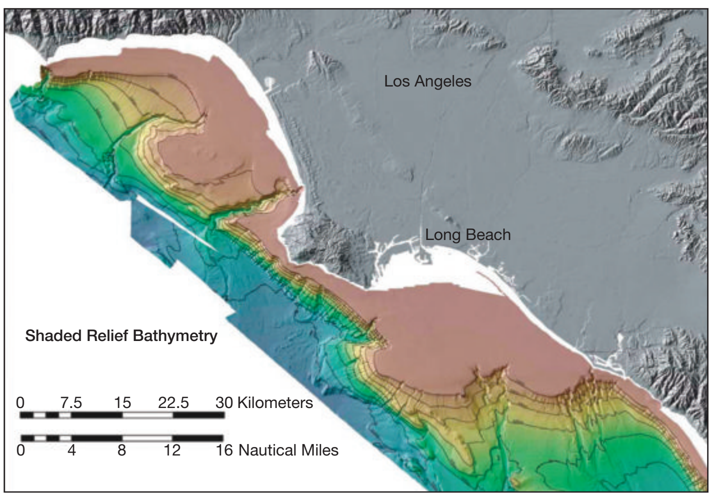


FIGURE 13.3 Color-enhanced perspective map of the seafloor and coastal landforms in the Los Angeles area of California. The ocean floor portion of this map (shown in color) was constructed from data collected using a high-resolution mapping system. (U.S. Geological Survey)

Viewing the Ocean Floor from Space

Another technological breakthrough that has led to an enhanced understanding of the seafloor involves measuring the shape of the surface of the global ocean from space. After compensating for waves, tides, currents, and atmospheric effects, it was discovered that the water's surface is not perfectly "flat." This is because gravity attracts water toward regions where massive seafloor features occur. Therefore, mountains and ridges produce elevated areas on the ocean surface, and, conversely, canyons and trenches cause slight depressions. Satellites equipped with *radar altimeters* are able to measure these subtle differences by bouncing microwaves off the sea surface (Figure 13.5). These devices can measure variations as small as 3 to 6 centimeters. Such data have added greatly to the knowledge of ocean-floor topography. Cross-checked with traditional sonar depth measurements, the data are used to produce detailed ocean-floor maps, such as the one shown in Figure 1.17 (p. 22).

Provinces of the Ocean Floor

Oceanographers studying the topography of the ocean floor have delineated three major units: *continental margins, deep-ocean basins,* and *oceanic (mid-ocean) ridges.* The map in Figure 13.6 outlines these provinces for the North Atlantic Ocean, and the profile at the bottom of the illustration shows the varied topography. Such profiles usually have their vertical dimension exaggerated many times—40 times in this case—to make topographic features more conspicuous. Vertical exaggeration, however, makes slopes shown in

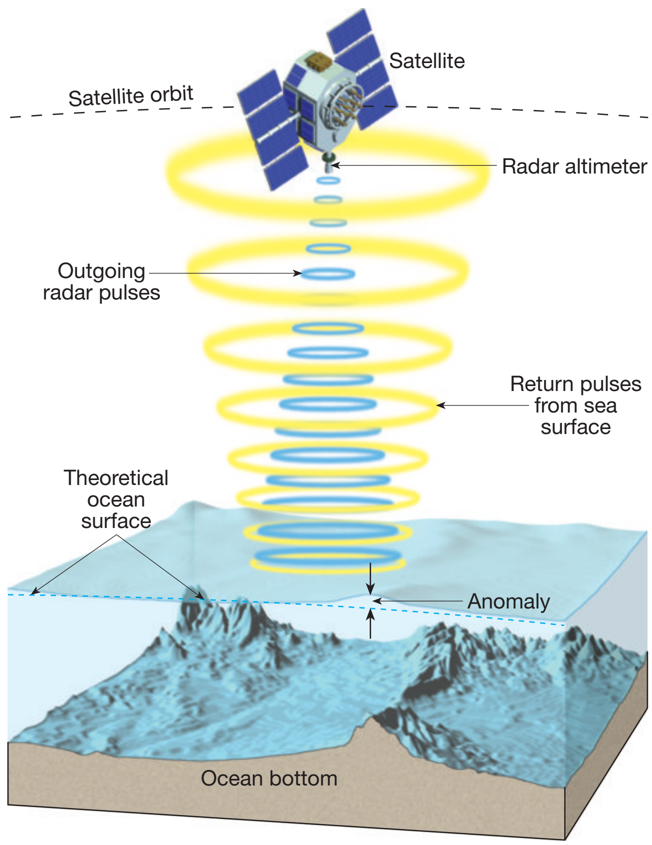


FIGURE 13.5 A satellite altimeter measures the variation in sea surface elevation, which is caused by gravitational attraction and mimics the shape of the seafloor. The sea surface anomaly is the difference between the measured and theoretical ocean surface.

seafloor profiles appear to be *much* steeper than they actually are.

Continental Margins

Two main types of **continental margins** have been identified—*passive* and *active.* Passive margins are found along most of the coastal areas that surround the Atlantic and Indian oceans, including the east coasts of North and South America, as well as the coastal areas of Europe and Africa. Passive margins are *not* situated along an active plate boundary and therefore experience very little volcanism and few earthquakes. Here, weathered materials eroded from the adjacent landmass accumulate to form a thick, broad wedge of relatively undisturbed sediments.

By contrast, active continental margins occur where oceanic lithosphere is being subducted beneath the edge of a continent. The result is a relatively narrow margin, consisting of highly deformed sediments that were scraped from the descending lithospheric slab. Active continental margins are common around the Pacific Rim, where they parallel deep-ocean trenches (see Box 13.1).

Passive Continental Margins

The features comprising a **passive continental margin** include the continental shelf, the continental slope, and the continental rise (Figure 13.7).

Continental Shelf The **continental shelf** is a gently sloping submerged surface extending from the shoreline toward the

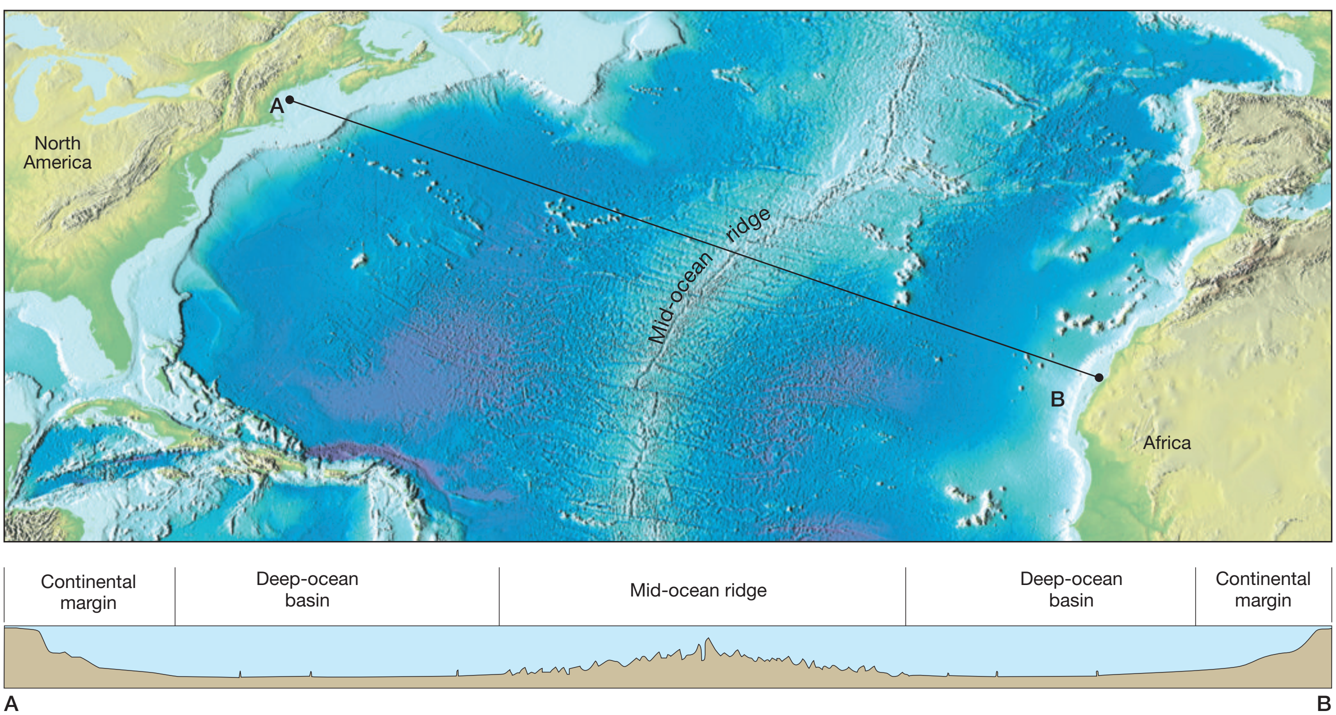


FIGURE 13.6 Major topographic divisions of the North Atlantic and a profile from New England to the coast of North Africa.

BOX 13.1 ▸ UNDERSTANDING EARTH

Susan DeBari—A Career in Geology

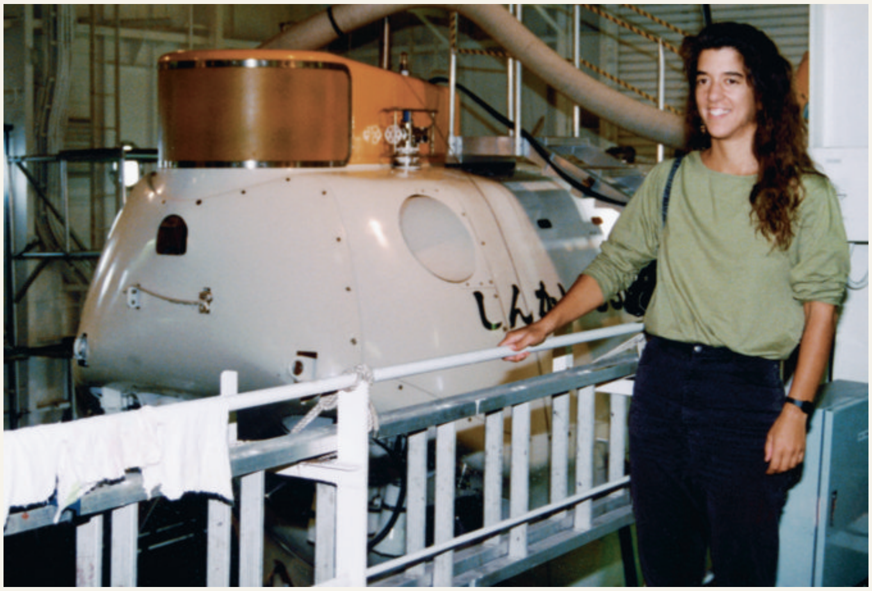


FIGURE 13.A Susan DeBari photographed with the Japanese submersible, *Shinkai 6500*, which she used to collect rock samples from the Izu Bonin trench. (Photo courtesy of Susan DeBari)

I discovered geology the summer I worked doing trail maintenance in the North Cascade mountains of Washington State. I had just finished my freshman year in college and had never before studied earth science. But a coworker (now my best friend) began to describe the geological features of the mountains that we were hiking in—the classic cone shape of Mount Baker volcano, the U-shaped glacial valleys, the advance of active glaciers, and other wonders. I was hooked and went back to college that fall with a geology passion that hasn't abated. As an undergraduate, I worked as a field assistant to a graduate student and did a senior thesis project on rocks from the Aleutian island arc. From that initial spark, island arcs have remained my top research interest, on through Ph.D. research at Stanford University, postdoctoral work at the University of Hawaii, and as a faculty member at San Jose State University and Western Washington University. Of most interest was the deep crust of arcs, the material that lies close to the Mohorovîĉiĉ discontinuity (fondly known as the Moho).

What kinds of processes are occurring down there at the base of the crust in island arcs? What is the source of magmas that make their way to the surface—the mantle, or the deep crust itself? How do these magmas interact with the crust as they make their way upward? What do these early magmas look like chemically? Are they very different from what is erupted at the surface?

Obviously, geologists cannot go down to the base of the crust (typically 20 to 40 kilometers beneath Earth's surface). So what they do is play a bit of a detective game. They must use rocks that are *now exposed at the surface* that were originally formed in the deep crust of an island arc. The rocks must have been brought to the surface rapidly along fault zones to preserve their original features. Thus, I can walk on rocks of the deep crust without really leaving the Earth's surface! There are a few places around the world where these rare rocks are exposed. Some of the places that I have worked include the Chugach Mountains of Alaska, the Sierras Pampeanas of Argentina, the Karakorum range in Pakistan, Vancouver Island's west coast, and the North Cascades of Washington. Fieldwork has most commonly involved hiking on foot, along with extensive use of mules and trucks.

I also went looking for exposed pieces of the deep crust of island arcs in a less obvious place, in one of the deepest oceanic trenches of the world, the Izu Bonin trench (Figure 13.A). Here I dove into the ocean in a submersible called the *Shinkai 6500* (pictured to my right in the background). The *Shinkai 6500* is a Japanese submersible that has the capability to dive to 6500 meters below the surface of the ocean (approximately 4 miles). My plan was to take rock samples from the wall of the trench at its deepest levels using the submersible's mechanical arm. Because preliminary data suggested that vast amounts of rock were exposed for several kilometers in a vertical sense, this could be a great way to sample the deep arc basement. I dove in the submersible three times, reaching a maximum depth of 6497 meters. Each dive lasted nine hours, spent in a space no bigger than the front seat of a Honda, shared with two of the Japanese pilots that controlled the submersible's movements. It was an exhilarating experience!

I am now on the faculty at Western Washington University, where I continue to do research on the deep roots of volcanic arcs, and get students involved as well. I am also involved in science education training for K–12 teachers, hoping to get young people motivated to ask questions about the fascinating world that surrounds them!

deep-ocean basin. Because it is underlain by continental crust, it is clearly a flooded extension of the continents.

The continental shelf varies greatly in width. Although almost nonexistent along some continents, the shelf extends seaward more than 1500 kilometers (930 miles) along others. On average, the continental shelf is about 80 kilometers (50 miles) wide and 130 meters (425 feet) deep at its seaward edge. The average inclination of the continental shelf is only about one-tenth of 1 degree, a drop of only about 2 meters per kilometer (10 feet per mile). The slope is so slight that it would appear to an observer to be a horizontal surface.

Although continental shelves represent only 7.5 percent of the total ocean area, they have economic and political significance because they contain important mineral deposits,

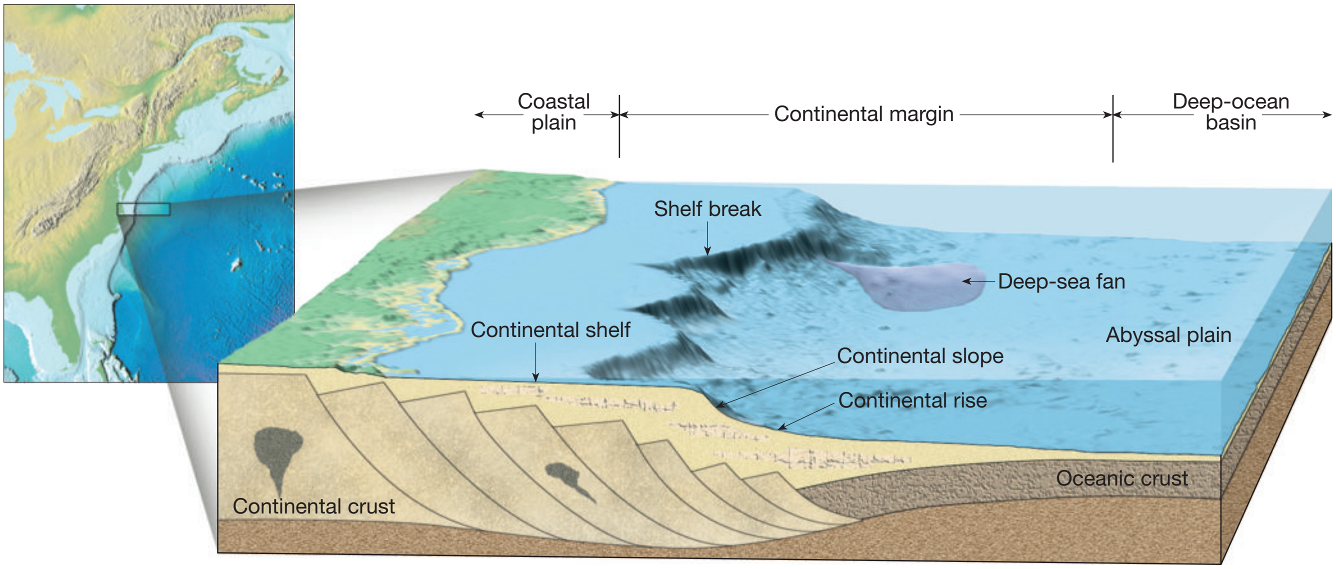


FIGURE 13.7 Schematic view showing the major features of a passive continental margin. Note that the slopes shown for the continental shelf and continental slope are greatly exaggerated. The continental shelf has an average slope of one-tenth of 1 degree, while the continental slope has an average slope of about 5 degrees.

including large reservoirs of oil and natural gas, as well as huge sand and gravel deposits. The waters of the continental shelf also contain many important fishing grounds, which are significant sources of food.

Even though the continental shelf is relatively featureless, some areas are mantled by extensive glacial deposits and thus are quite rugged. In addition, some continental shelves are dissected by large valleys running from the coastline into deeper waters. Many of these *shelf valleys* are the seaward extensions of river valleys on the adjacent landmass. Such valleys appear to have been excavated during the Pleistocene epoch (Ice Age). During this time great quantities of water were stored in vast ice sheets on the continents. This caused sea level to drop by 100 meters (330 feet) or more, exposing large areas of the continental shelves. Because of this drop in sea level, rivers extended their courses, and land-dwelling plants and animals inhabited the newly exposed portions of the continents. Dredging off the coast of North America has retrieved the ancient remains of numerous land dwellers, including mammoths, mastodons, and horses, adding to the evidence that portions of the continental shelves were once above sea level.

Most passive continental shelves, such as those along the East Coast of the United States, consist of shallow-water deposits that can reach several kilometers in thickness. Such deposits have led researchers to conclude that these thick accumulations of sediment are produced along a gradually subsiding continental margin.

Continental Slope Marking the seaward edge of the continental shelf is the **continental slope,** a relatively steep structure (as compared with the shelf) that marks the boundary between continental crust and oceanic crust (Figure 13.7). Although the inclination of the continental slope varies greatly from place to place, it averages about 5 degrees and in places may exceed 25 degrees. Further, the continental slope is a relatively narrow feature, averaging only about 20 kilometers (12 miles) in width.

Continental Rise In regions where trenches do not exist, the steep continental slope merges into a more gradual incline known as the **continental rise.** Here the slope drops to about one-third a degree, or about 6 meters per kilometer (32 feet per mile). Whereas the width of the continental slope averages about 20 kilometers (12 miles), the continental rise may extend for hundreds of kilometers into the deep-ocean basin.

The continental rise consists of a thick accumulation of sediment that moved downslope from the continental shelf to the deep-ocean floor. The sediments are delivered to the base of the continental slope by *turbidity currents* that periodically flow down submarine canyons. When these muddy currents emerge from the mouth of a canyon onto the relatively flat ocean floor, they deposit sediment that forms a **deep-sea fan** (Figure 13.7). As fans from adjacent submarine canyons grow, they merge laterally with one another to produce a continuous covering of sediment at the base of the continental slope forming the continental rise.

Active Continental Margins

Along some coasts the continental slope descends abruptly into a deep-ocean trench. In this situation, the landward wall of the trench and the continental slope are essentially the same feature. In such locations, the continental shelf is very narrow, if it exists at all.

Active continental margins are located primarily around the Pacific Ocean in areas where oceanic lithosphere is being subducted beneath the leading edge of a

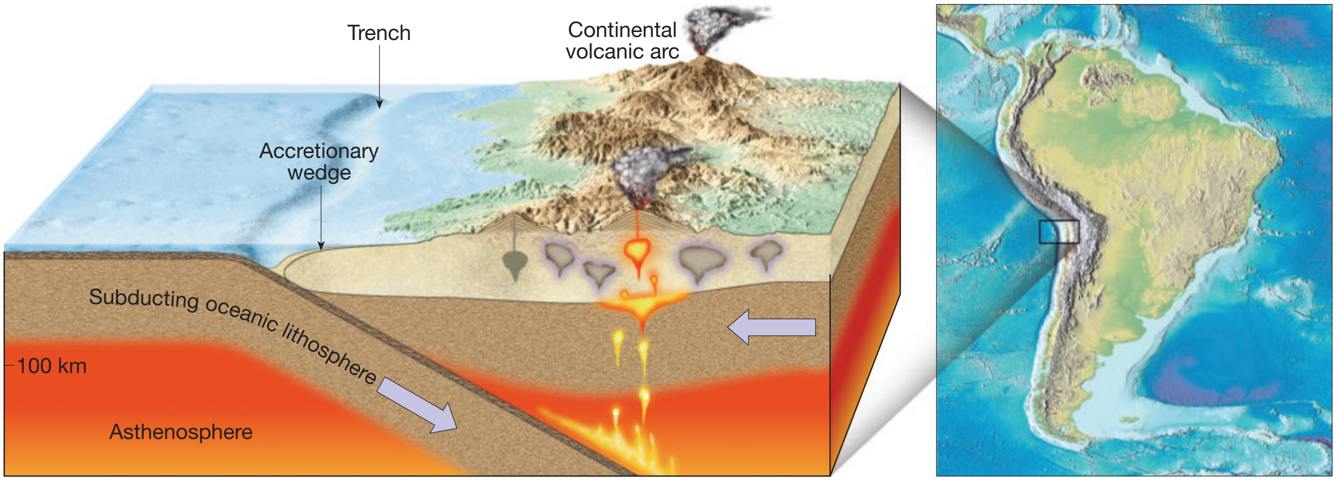


FIGURE 13.8 Active continental margin. Here sediments from the ocean floor are scraped from the descending plate and added to the continental crust as an accretionary wedge.

continent (Figure 13.8). Here sediments from the ocean floor and pieces of oceanic crust are scraped from the descending oceanic plate and plastered against the edge of the overriding continent. This chaotic accumulation of deformed sediment and scraps of oceanic crust is called an **accretionary** (*ad* = toward, *crescere* = to grow) **wedge.** Prolonged plate subduction, along with the accretion of sediments on the landward side of the trench, can produce a large accumulation of sediments along a continental margin. A large accretionary wedge, for example, is found along the northern coast of Japan's Honshu Island.

Some subduction zones have little or no accumulation of sediments, indicating that ocean sediments are being carried into the mantle with the subducting plate. These tend to be regions where old oceanic lithosphere is subducting nearly vertically into the mantle. In these locations the continental margin is very narrow, as the trench may lie a mere 50 kilometers (31 miles) offshore.

Features of Deep-Ocean Basins

Between the continental margin and the oceanic ridge lies the **deep-ocean basin** (see Figure 13.6). The size of this region—almost 30 percent of Earth's surface—is roughly comparable to the percentage of land above sea level. This region includes remarkably flat areas known as *abyssal plains;* tall volcanic peaks called *seamounts* and *guyots; deep-ocean trenches,* which are extremely deep linear depressions in the ocean floor; and large flood basalt provinces called *oceanic plateaus.*

Deep-Ocean Trenches

Deep-ocean trenches are long, relatively narrow creases in the seafloor that form the deepest parts of the ocean (Table 13.1). Most trenches are located along the margins of the Pacific Ocean (Figure 13.9), where many exceed 10,000 meters (33,000 feet) in depth. A portion of one trench—the Challenger Deep in the Mariana Trench—has been measured at 11,022 meters (36,163 feet) below sea level, making it the deepest known part of the world ocean. Only two trenches are located in the Atlantic—the Puerto Rico Trench adjacent to the Lesser Antilles arc and the South Sandwich Trench.

Although deep-ocean trenches represent only a small portion of the area of the ocean floor, they are nevertheless significant geologic features. Trenches are sites of plate convergence where lithospheric plates subduct and plunge back into the mantle. In addition to earthquakes being created as one plate "scrapes" against another, volcanic activity is also associated with these regions. Thus, trenches are often paralleled by an arc-shaped row of active volcanoes called a *volcanic island arc.* Furthermore, *continental volcanic arcs,* such as those making up portions of the Andes and Cascades, are located parallel to trenches that lie adjacent to continental margins. The large number of trenches and associated volcanic activity along the margins of the Pacific Ocean explains why the region is known as the *Ring of Fire.*

TABLE 13.1 Dimensions of Some Deep-Ocean Trenches

Trench	Depth (kilometers)	Average Width (kilometers)	Length (kilometers)
Aleutian	7.7	50	3700
Japan	8.4	100	800
Java	7.5	80	4500
Kurile–Kamchatka	10.5	120	2200
Mariana	11.0	70	2550
Central America	6.7	40	2800
Peru–Chile	8.1	100	5900
Philippine	10.5	60	1400
Puerto Rico	8.4	120	1550
South Sandwich	8.4	90	1450
Tonga	10.8	55	1400

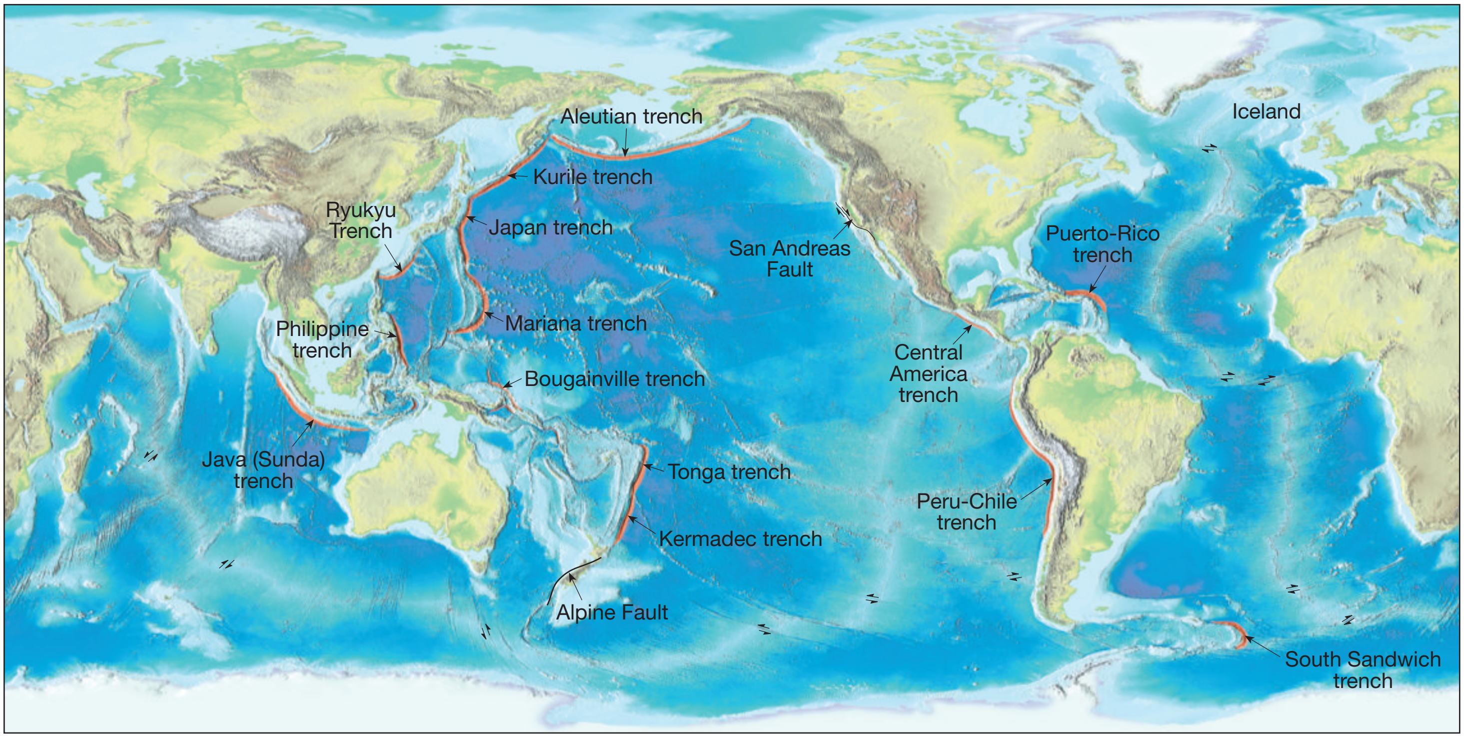


FIGURE 13.9 Distribution of the world's deep-ocean trenches.

Abyssal Plains

Abyssal (*a* = without, *byssus* = bottom) **plains** are deep, incredibly flat features; in fact, these regions are likely the most level places on Earth. The abyssal plain found off the coast of Argentina, for example, has less than 3 meters (10 feet) of relief over a distance exceeding 1300 kilometers (800 miles). The monotonous topography of abyssal plains is occasionally interrupted by the protruding summit of a partially buried volcanic peak.

Using *seismic profilers* (instruments that generate signals designed to penetrate far below the ocean floor), researchers have determined that abyssal plains owe their relatively featureless topography to thick accumulations of sediment that have buried an otherwise rugged ocean floor (see Figure 13.4). The nature of the sediment indicates that these plains consist primarily of fine sediments transported far out to sea by turbidity currents, deposits that have precipitated out of seawater, and shells and skeletons of microscopic marine organisms.

Abyssal plains are found in all oceans. However, the Atlantic Ocean has the most extensive abyssal plains because it has few trenches to act as traps for sediment carried down the continental slope.

Students Sometimes Ask . . .

Have humans ever explored the deepest ocean trenches? Could anything live there?

Humans have indeed visited the deepest part of the oceans—where there is crushing high pressure, complete darkness, and near-freezing water temperatures—more than 45 years ago! In January 1960, U.S. Navy Lt. Don Walsh and explorer Jacques Piccard descended to the bottom of the Challenger Deep region of the Mariana Trench in the deep-diving bathyscaphe *Trieste.* At 9906 meters (32,500 feet), the men heard a loud cracking sound that shook the cabin. They were unable to see that a 7.6-centimeter (3-inch) Plexiglas viewing port had cracked (miraculously, it held for the rest of the dive). More than five hours after leaving the surface, they reached the bottom at 10,912 meters (35,800 feet)—a record depth of human descent that has not been broken since. They did see some life forms that are adapted to life in the deep: a small flatfish, a shrimp, and some jellyfish.

Seamounts, Guyots, and Oceanic Plateaus

Dotting the ocean floor are submarine volcanoes called **seamounts,** which may rise hundreds of meters above the surrounding topography. It is estimated that more than a million of these features exist. Some grow large enough to become oceanic islands, but these are rare. Most do not have a sufficiently long eruptive history to build a structure above sea level. Although these conical peaks are found on the floors of all the oceans, the greatest number have been identified in the Pacific. Furthermore, seamounts often form linear chains, or in some cases a more continuous volcanic ridge, not to be confused with mid-ocean ridges.

BOX 13.2 ▸ UNDERSTANDING EARTH

Explaining Coral Atolls—Darwin's Hypothesis

Coral *atolls* are ring-shaped structures that often extend several thousand meters below sea level (Figure 13.B). What causes atolls to form, and how do they attain such a great thickness?

Corals are colonial animals about the size of an ant that feed with stinging tentacles and are related to jellyfish. Most corals protect themselves by creating a hard external skeleton made of calcium carbonate. Where corals reproduce and grow over many centuries their skeletons fuse into large structures called *coral reefs.* Other corals—as well as sponges and algae—begin to attach to the reef, enlarging it further. Eventually fishes, sea slugs, octopus, and other organisms are attracted to these diverse and productive habitats.

FIGURE 13.B An aerial view of Tetiaroa Atoll in the Pacific. The light blue waters of the relatively shallow lagoon contrast with the dark blue color of the deep ocean surrounding the atoll. (Photo by Douglas Peebles Photography)

Corals require specific environmental conditions to grow. For example, reef-building corals grow best in waters with an average annual temperature of about 24°C (75°F). They cannot survive prolonged exposure to temperatures below 18°C (64°F) or above 30°C (86°F). In addition, reef-builders require an attachment site (usually other corals) and clear, sunlit water. Consequently, the limiting depth of most active reef growth is only about 45 meters (150 feet).

The restricted environmental conditions required for coral growth create an interesting paradox: How can corals—which require warm, shallow, sunlit water no deeper than a few dozen meters to live—create thick structures such as coral atolls that extend into deep water?

The naturalist Charles Darwin was one of the first to formulate a hypothesis on the origin of atolls. From 1831 to 1836 he sailed aboard the British ship HMS *Beagle* during its famous circumnavigation of the globe. In various places that Darwin visited, he noticed a progression of stages in coral reef development from (1) a *fringing reef* along the margins of a volcano to (2) a *barrier reef* with a volcano in the middle to (3) an *atoll,* which consists of a continuous or broken ring of coral reef surrounded by a central lagoon (Figure 13.C). The essence of Darwin's hypothesis was that as a volcanic island slowly sinks, the corals continue to build the reef complex upward.

Some, like the Hawaiian Island–Emperor Seamount chain in the Pacific, which stretches from the Hawaiian Islands to the Aleutian trench, form over a volcanic hot spot in association with a mantle plume (see Figure 2.28, p. 63). Others are born near oceanic ridges. If the volcano grows large enough before being carried from the magma source by plate movement, the structure may emerge as an island. Examples in the Atlantic include the Azores, Ascension, Tristan da Cunha, and St. Helena.

During the time they exist as islands, some of these volcanic structures are lowered to near sea level by the forces of weathering and erosion. In addition, islands will gradually sink and disappear below the water surface as the moving plate slowly carries them away from the elevated oceanic ridge or hot spot where they originated (see Box 13.2). Submerged, flat-topped seamounts which formed in this manner are called **guyots** or **tablemounts.***

*The term *guyot* is named after Princeton University's first geology professor. It is pronounced "GEE-oh" with a hard *g* as in "give."

Mantle plumes have also generated several large **oceanic plateaus,** which resemble the flood basalt provinces found on the continents. Examples of these extensive volcanic structures include the Ontong Java and Caribbean plateaus, which formed from vast outpourings of fluid basaltic lavas onto the ocean floor (Figure 13.10). Hence, oceanic plateaus are composed mostly of basalts and ultramafic rocks that in some cases exceed 30 kilometers in thickness.

Anatomy of the Oceanic Ridge

GEODe EARTH Divergent Boundaries
▸ Oceanic Ridges and Seafloor Spreading

Along well-developed divergent plate boundaries, the seafloor is elevated, forming a broad linear swell called the **oceanic ridge,** or **mid-ocean ridge.** Our knowledge of the oceanic ridge system comes from soundings taken of the ocean floor, core samples obtained from deep-sea drilling, visual inspection using deep-diving submersibles

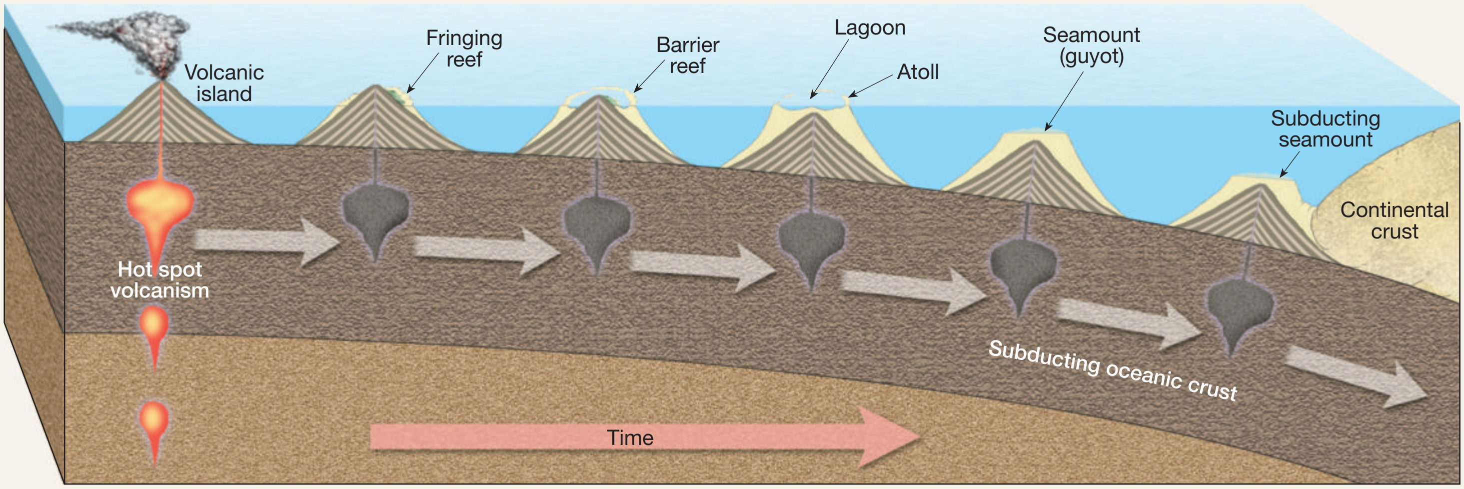


FIGURE 13.C Formation of a coral atoll due to the gradual sinking of oceanic crust and upward growth of the coral reef. A fringing coral reef forms around an active volcanic island. As the volcanic island moves away from the region of hotspot activity it sinks, and the fringing reef gradually becomes a barrier reef. Eventually, the volcano is completely submerged and an atoll remains.

Darwin's hypothesis explained how coral reefs, which are restricted to shallow water, can build structures that now exist in much deeper water. During Darwin's time, however, there was no plausible mechanism to account for how an island might sink.

Today, plate tectonics helps explain how a volcanic island can become extinct and sink to great depths over long periods of time. Volcanic islands often form over a relatively stationary mantle plume, which causes the lithosphere to be buoyantly uplifted. Over a span of millions of years, these volcanic islands become inactive and gradually sink as the moving plate carries them away from the region of hotspot volcanism (Figure 13.C).

Furthermore, drilling through atolls has revealed that volcanic rock does indeed underlie the oldest (and deepest) coral reef structures, confirming Darwin's hypothesis. Thus, atolls owe their existence to the gradual sinking of volcanic islands containing coral reefs that build upward through time.

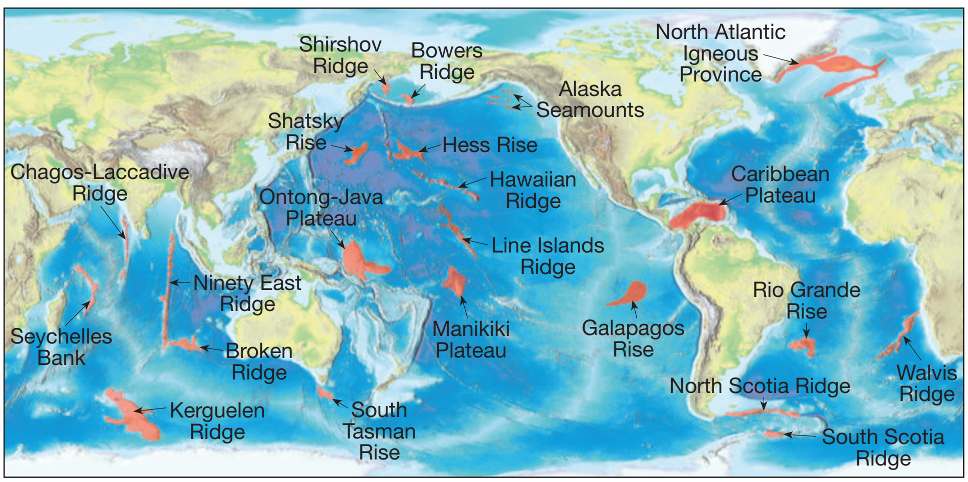


FIGURE 13.10 Distribution of oceanic plateaus, hot spot tracks, and other submerged crustal fragments.

(Figure 13.11), and even firsthand inspection of slices of ocean floor that have been displaced onto dry land along convergent plate boundaries. An elevated position, extensive faulting and associated earthquakes, high heat flow, and numerous volcanic structures characterize the oceanic ridge.

The interconnected oceanic ridge system is the longest topographic feature on Earth's surface, exceeding 70,000 kilometers (43,000 miles) in length. Representing more than 20 percent of Earth's surface, the oceanic ridge winds through all major oceans in a manner similar to the seam on a baseball (Figure 13.12). The crest of this linear structure typically stands 2 to 3 kilometers above the adjacent deep-ocean basins and marks the plate margins where new oceanic crust is created.

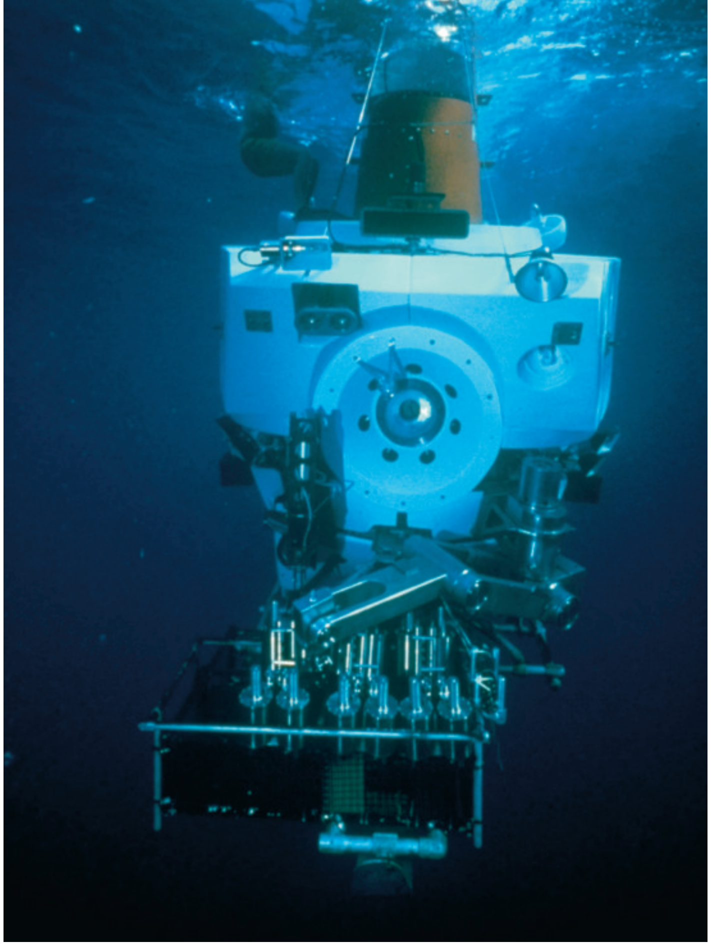


FIGURE 13.11 The deep-diving submersible *Alvin* is 7.6 meters long, weighs 16 tons, has a cruising speed of 1 knot, and can reach depths as great as 4000 meters. A pilot and two scientific observers are along during a normal 6- to 10-hour dive. (Courtesy of Rod Catanach/Woods Hole Oceanographic Institution)

Notice in Figure 13.12 that large sections of the oceanic ridge system have been named based on their locations within the various ocean basins. Some ridges run along the middle of ocean basins, where they are called *mid-ocean* ridges. This holds true for the Mid-Atlantic Ridge, which is positioned in the middle of the Atlantic, roughly paralleling the margins of the continents on either side. This is also true for the Mid-Indian Ridge. However, the East Pacific Rise is not a "mid-ocean" feature. Rather, as its name implies, it is located in the eastern Pacific, far from the center of the ocean. At its northern end there are two branches—one that points toward Central America and another that curves toward South America (Figure 13.12).

The term *ridge* may be misleading, because these features are not narrow and steep as the term implies, but have widths of from 1000 to 4000 kilometers and the appearance of a broad, elongated swell that exhibits various degrees of ruggedness. Furthermore, the ridge system is broken into segments that range from a few tens to hundreds of kilometers in length. Although each segment is offset from the adjacent segment, they are generally connected, one to the next, by a transform fault.

Oceanic ridges are as high as some mountains found on the continents; however, the similarity ends there. Whereas most continental mountains form when compressional forces fold and metamorphose thick sequences of sedimentary rocks along convergent plate boundaries, oceanic ridges form where tensional forces fracture and pull the ocean crust apart. The oceanic ridge consists of layers and piles of newly formed mafic rocks that have

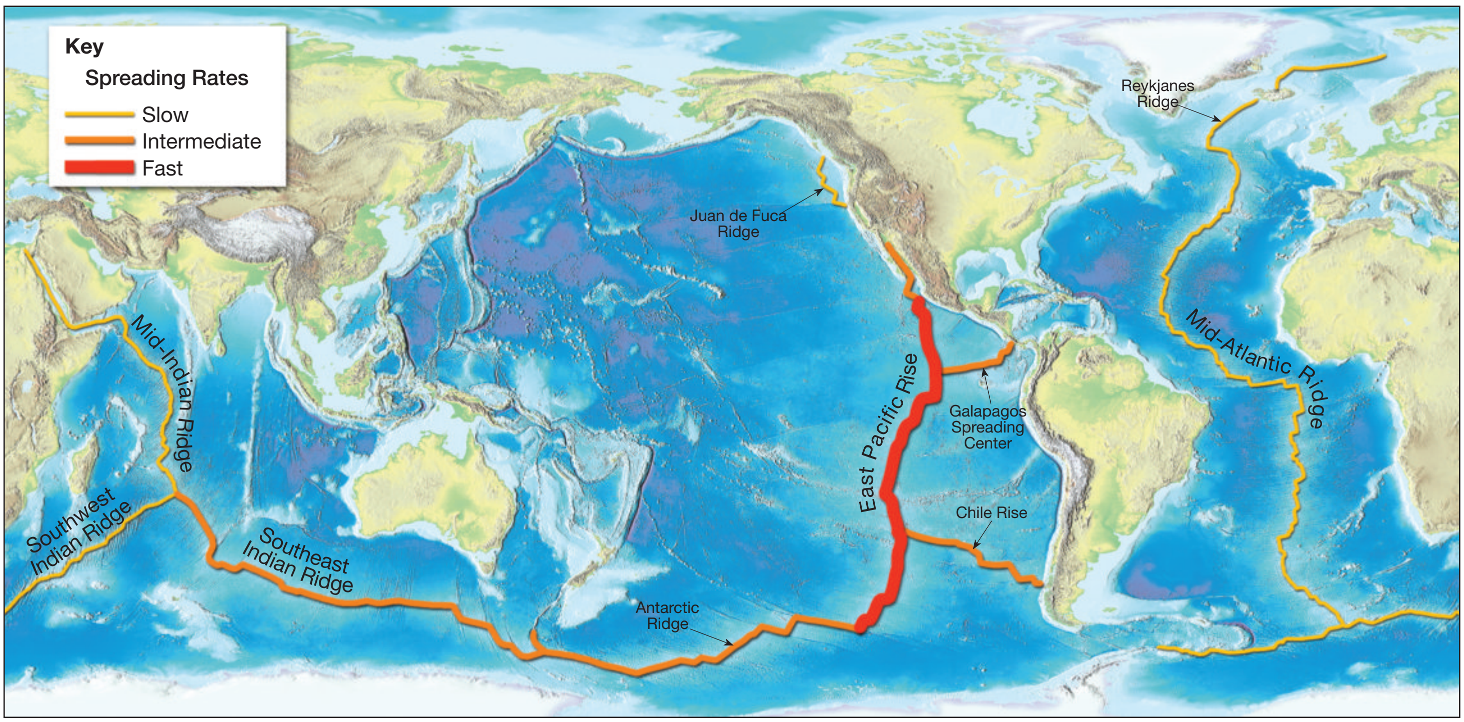


FIGURE 13.12 Distribution of the oceanic ridge system. The map shows ridge segments that exhibit slow, intermediate, and fast spreading rates.

been faulted into elongated blocks that are buoyantly uplifted.

Along the axis of some segments of the oceanic ridge system are deep, down-faulted structures called **rift valleys** (Figure 13.13). The name *rift valley* has been applied to these features because they are strikingly similar to the continental rift valleys that comprise the East African Rift (see Figure 13.20). Some rift valleys, including many along the Mid-Atlantic Ridge, exceed 30 kilometers in width and have walls that tower 2000 meters above the valley floor. This makes them comparable to the deepest and widest part of Arizona's Grand Canyon.

Oceanic Ridges and Seafloor Spreading

GEODe EARTH Divergent Boundaries
▸ Oceanic Ridges and Seafloor Spreading

The greatest volume of magma (more than 60 percent of Earth's total yearly output) is produced along the oceanic ridge system in association with seafloor spreading. As the plates diverge, fractures are created in the oceanic crust that fill with molten rock that wells up from the hot asthenosphere below. This molten material slowly cools to solid rock, producing new slivers of seafloor. This process occurs again and again, generating new lithosphere that moves from the ridge crest in a conveyor belt fashion.

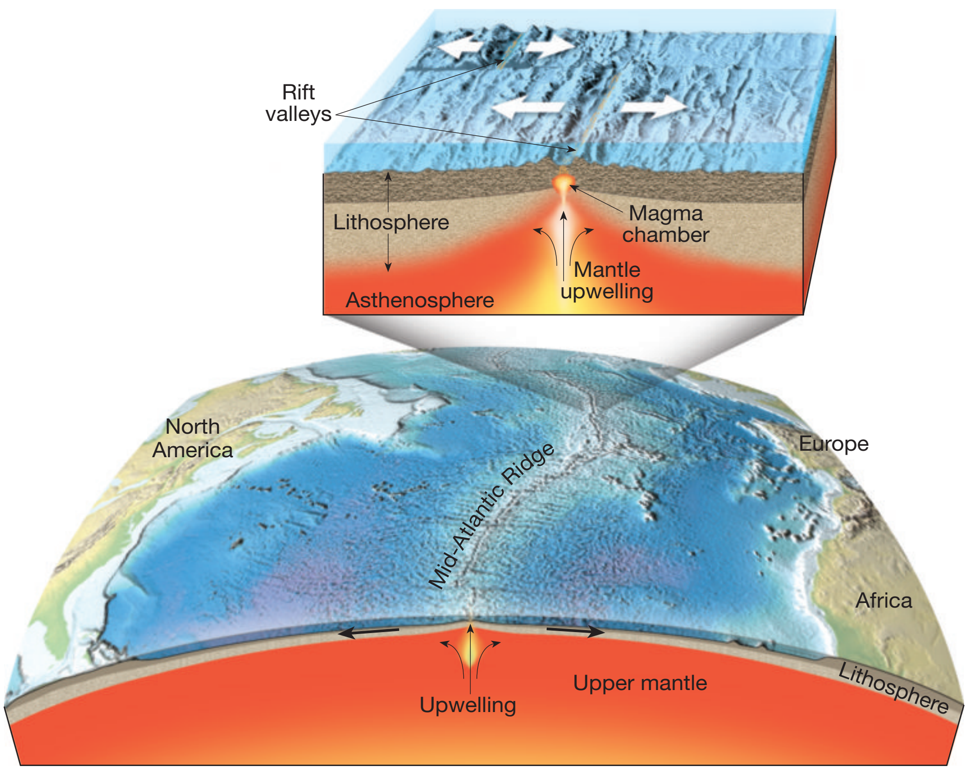


FIGURE 13.13 The axis of some segments of the oceanic ridge system contains deep downfaulted structures called *rift valleys.* Some may exceed 30 kilometers in width and 2000 meters in depth.

Seafloor Spreading

Harry Hess of Princeton University formulated the concept of seafloor spreading in the early 1960s. Later, geologists were able to verify Hess's view that seafloor spreading occurs along relatively narrow areas, located at the crests of oceanic ridges. Here, below the ridge axis where the lithospheric plates separate, solid hot mantle rocks rise upward to replace the material that has shifted horizontally. Recall from Chapter 4 that as rock rises, it experiences a decrease in confining pressure and may undergo melting without the addition of heat. This process, called *decompression melting,* is how magma is generated along the ridge axis.

Partial melting of mantle rock produces basaltic magma having a composition that is surprisingly consistent along the entire length of the ridge system. This newly formed magma separates from the mantle rock from which it was derived, and rises toward the surface in the form of teardrop-shaped blobs. Although most of this magma is thought to collect in elongated reservoirs (magma chambers) located just beneath the ridge crest, about 10 percent eventually migrates upward along fissures to erupt as lava flows on the ocean floor (Figure 13.13). This activity continuously adds new basaltic rock to the plate margins, temporarily welding them together, only to be broken as spreading continues. Along some ridges, outpourings of bulbous lavas build submerged shield volcanoes (seamounts) as well as elongated lava ridges. At other locations, more voluminous lava flows create a relatively subdued topography.

During seafloor spreading, the magma that is injected into newly developed fractures forms dikes that cool from their outer borders inward toward their centers. Because the warm interiors of these newly formed dikes are weak, continued spreading produces new fractures that tend to split these young rocks roughly in half. As a result, new material is added equally to the two diverging plates. Consequently, new ocean floor grows symmetrically on each side of the centrally located ridge crest. Indeed, the ridge systems of the Atlantic and Indian oceans are located near the middle of these water bodies. However, the East Pacific Rise is situated far from the center of the Pacific Ocean. Despite uniform spreading along the East Pacific Rise, much of the Pacific Basin that once lay east of this divergent boundary has been overridden by the westward migration of the American plates.

When Harry Hess first proposed the concept of seafloor spreading, upwelling in the mantle was thought to be one of the driving forces for plate motions. Geologists have since discovered that upwelling along the oceanic ridge is a *passive process*. Stated another way, mantle upwelling occurs because "space" is created as oceanic lithosphere moves horizontally away from the ridge axis.

Why Are Oceanic Ridges Elevated?

The primary reason for the elevated position of a ridge system is the fact that newly created oceanic lithosphere is hot, occupies more volume, and is therefore less dense than cooler rocks of the deep-ocean basin. As the newly formed basaltic crust travels away from the ridge crest, it is cooled from above as seawater circulates through the pore spaces and fractures in the rock. It also cools because it is moving away from the zone of upwelling, which is the main source of heat. As a result, the lithosphere gradually cools, contracts, and becomes more dense. This thermal contraction accounts for the greater ocean depths that exist away from the ridge. It takes almost 80 million years before cooling and contraction cease completely. By this time, rock that was once part of an elevated ocean-ridge system is located in the deep-ocean basin, where it may be covered by relatively thick accumulations of sediment.

As the lithosphere is displaced away from the ridge crest, cooling also causes a gradual increase in lithospheric thickness. This occurs because the boundary between the lithosphere and asthenosphere is temperature dependent. Recall that the lithosphere is Earth's cool, rigid outer layer, whereas the asthenosphere is a comparatively hot and weak zone. As material in the upper mantle ages (cools), it becomes rigid. Thus, the upper portion of the asthenosphere is converted to lithosphere simply by cooling. Newly formed oceanic lithosphere will continue to thicken for about 80 million years. Thereafter its thickness remains relatively constant until it is subducted.

Spreading Rates and Ridge Topography

When various segments of the oceanic ridge system were studied in detail, some topographic differences came to light. Many of these differences appear to be controlled by spreading rates. One of the main factors controlled by spreading rates is the amount of magma generated at a rift zone. At fast spreading centers, divergence occurs at a greater rate than at slow spreading centers, resulting in more magma welling up from the mantle. As a result, the magma chambers located below fast spreading centers tend to be larger and more permanent features than those associated with slower spreading centers. Further, spreading along fast spreading centers appears to be a relatively continuous process where rifting and upwelling is occurring along the entire length of the ridge axis. By contrast, rifting at slow spreading centers appears to be more episodic, where segments of the ridge may remain dormant for extended intervals.

At slow spreading rates of 1 to 5 centimeters per year, such as occur at the Mid-Atlantic and Mid-Indian ridges, a prominent rift valley is present along much of the ridge crest and the topography is quite rugged (Figure 13.14A). Recall that these rift valleys can exceed 30 kilometers in width and 2000 meters in depth. Here the upward displacement of large, buoyant slabs of oceanic crust along nearly vertical faults produces the steep walls of the rift valley. In addition, at slow spreading centers volcanism produces numerous volcanic cones within the rift valley which enhance the rugged topography of the ridge crest.

Along the Galápagos ridge an intermediate spreading rate of 5 to 9 centimeters per year is the norm. In settings such as this the rift valleys that develop are shallow, often less than 200 meters deep. In addition, their topography tends to be subdued compared to ridges that exhibit slower spreading rates.

At faster spreading rates (greater than 9 centimeters per year), such as occur along much of the East Pacific Rise, rift valleys are generally absent (Figure 13.14B). Instead, the ridge axis is higher than the ocean floor on either side. Such areas consist of *swells*—volcanic extrusions that tend to overlap or even produce relatively narrow structures (Figure 13.15). In other places along fast spreading centers, sheets of fluid lava have produced areas of relatively subdued topography. Because the depth of the ocean depends on the age of the seafloor, ridge segments that exhibit faster

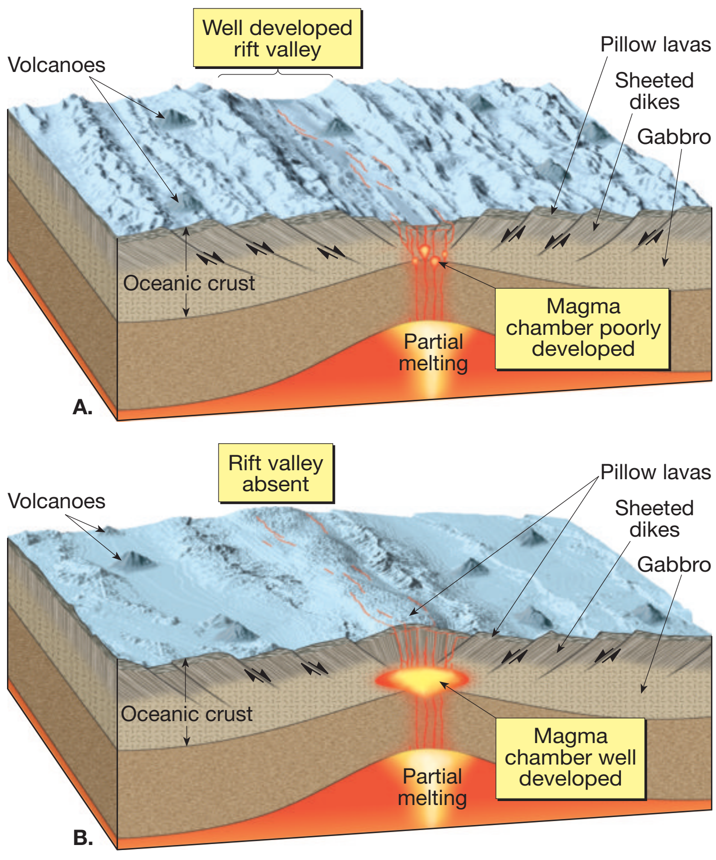


FIGURE 13.14 Topography of the crest of an oceanic ridge. **A.** At slow spreading rates a prominent rift valley develops along the ridge crest, and the topography is typically rugged. **B.** Along fast spreading centers no median rift valleys develop, and the topography is comparatively smooth.

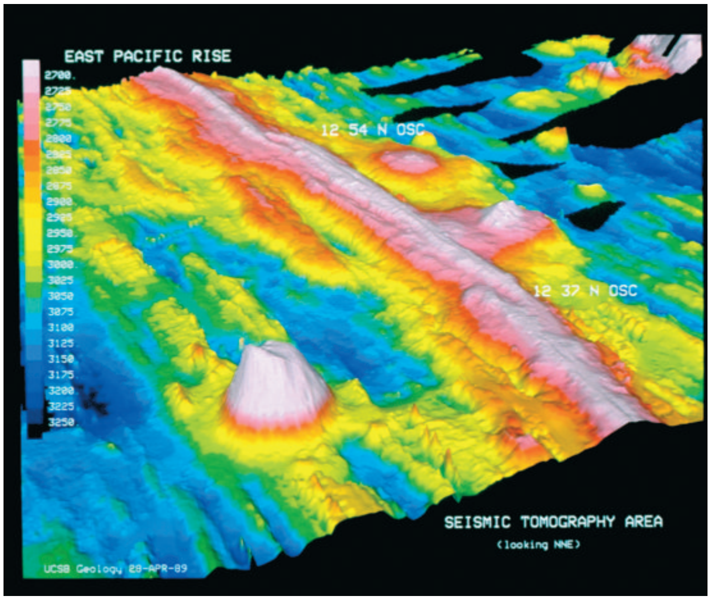


FIGURE 13.15 False-color sonar image of a segment of the East Pacific Rise. The linear pink area is the swell that formed above the ridge axis. Note also the large volcanic cone in the lower left portion of the image. (Courtesy of S. P. Miller)

spreading rates tend to have more gradual profiles than ridges that have slower spreading rates (Figure 13.16). Because of these differences in topography, the gently sloping, less rugged portions of the oceanic ridges are called *rises.*

The Nature of Oceanic Crust

One of the most interesting aspects of the oceanic crust is that its thickness and structure is remarkably consistent throughout the entire ocean basin. Seismic soundings indicate that it

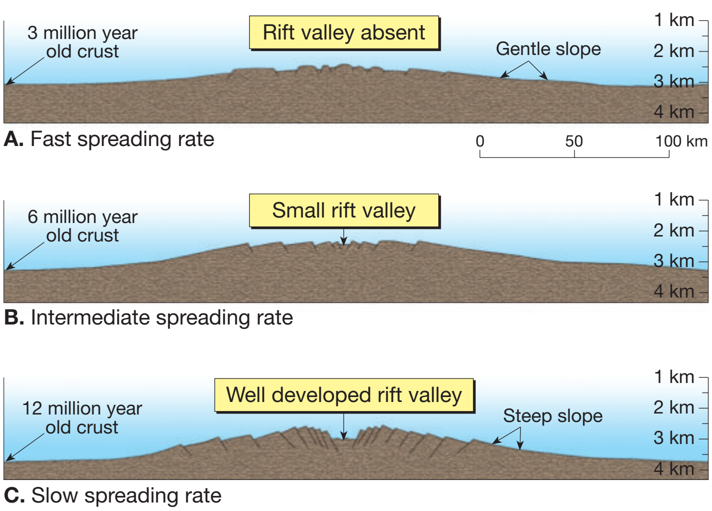


FIGURE 13.16 Schematic of ridge segments that exhibit fast, intermediate, and slow spreading rates. Fast spreading centers have gentle slopes and lack a rift valley. By contrast, ridges that have slow spreading rates have well-developed rift valleys and steep flanks. (The slopes of all these profiles are greatly exaggerated.)

averages only about 7 kilometers (5 miles) in thickness. Furthermore, it is composed almost entirely of mafic (basaltic) rocks that are underlain by a layer of the ultramafic rock peridotite, which forms the lithospheric mantle.

Although most oceanic crust forms out of view, far below sea level, geologists have been able to examine the structure of the ocean floor firsthand. In such locations as Newfoundland, Cyprus, Oman, and California, slivers of oceanic crust have been thrust high above sea level. From these exposures, researchers conclude that the ocean crust consists of four distinct layers (Figure 13.17):

- Layer 1: The upper layer is comprised of a sequence of unconsolidated sediments. Sediments are very thin near the axes of oceanic ridges, but may be several kilometers thick next to continents.
- Layer 2: Below the layer of sediments is a rock unit composed mainly of basaltic lavas that contain abundant pillowlike structures called *pillow basalts.*
- Layer 3: The middle rocky layer is made up of numerous interconnected dikes having a nearly vertical orientation, called the *sheeted dike complex.* These dikes are former pathways where magma rose to feed lava flows on the ocean floor.
- Layer 4: The lower unit is made up mainly of gabbro, the coarse-grained equivalent of basalt, which crystallized in a magma chamber below the ridge axis.

This sequence of rocks composing the oceanic crust is called an **ophiolite complex** (Figure 13.17). From studies of various ophiolite complexes around the globe and related data, geologists have pieced together a scenario for the formation of the ocean floor.

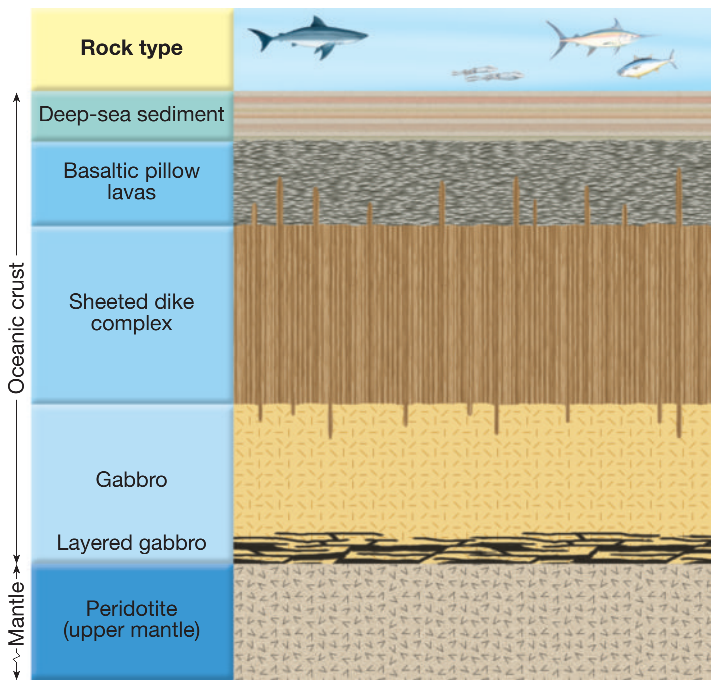


FIGURE 13.17 Rock types associated with a typical section of oceanic crust based on data obtained from ophiolite complexes and seismic studies.

How Does Oceanic Crust Form?

Recall that the basaltic magma needed to create new ocean crust originates from partially melted mantle rock (peridotite). Being partially molten and less dense than the surrounding solid rock, the magma gradually moves upward, where it enters a magma chamber that is thought to be less than 10 kilometers wide and located only 1 or 2 kilometers below the ridge crest. Seismic studies conducted along the East Pacific Rise have identified magma chambers along 60 percent of the ridge. Hence, these structures appear to be relatively permanent features, at least along fast spreading centers. However, along slow spreading ridges, where the rate of magma production is less, magma chambers are small and appear to form intermittently.

As seafloor spreading proceeds, numerous vertical fractures develop in the ocean crust that lies above these magma chambers. Molten rock is injected into these fissures, where some of it cools and solidifies to form dikes. New dikes intrude older dikes, which are still warm and weak, to form the **sheeted dike complex.** This portion of the oceanic crust is usually 1 to 2 kilometers thick.

Roughly 10 percent of the magma entering the reservoirs eventually erupts on the ocean floor. Because the surface of a submarine lava flow is chilled quickly by seawater, it rarely travels more than a few kilometers before completely solidifying. The forward motion occurs as lava accumulates behind the congealed margin and then breaks through. This process occurs over and over, as molten basalt is extruded—like toothpaste out of a tightly squeezed tube (Figure 13.18). The result is tube-shaped protuberances resembling large bed pillows stacked one atop the other, hence the name **pillow basalts** (Figure 13.19). In some settings, pillow lavas may build into volcano-size mounds that resemble shield volcanoes, whereas in others they form elongated ridges tens of kilometers long (Figure 13.18). These structures will eventually be cut off from their supply of magma as they are carried away from the ridge crest by seafloor spreading.

FIGURE 13.18 A photograph taken from the *Alvin* during Project FAMOUS shows lava extrusions in the rift valley of the Mid-Atlantic Ridge. Large toothpastelike extrusions such as this were common features. A mechanical arm is sampling an adjacent blisterlike extrusion. (Photo courtesy of Woods Hole Oceanographic Institution)

FIGURE 13.19 Pillow lava exposed along a sea cliff, Cape Wanbrow, New Zealand. Notice that each "pillow" shows an outer, rapidly cooled, dark glassy layer enclosing a dark gray basalt interior. (Photo by G. R. Roberts)

The lowest unit of the ocean crust develops from crystallization within the central magma chamber itself. The first minerals to crystallize are olivine, pyroxene, and occasionally chromite (chromium oxide), which fall through the magma to form a layered zone near the floor of the reservoir. The remaining magma tends to cool along the walls of the chamber and forms massive amounts of coarse-grained gabbro. This unit makes up the bulk of the oceanic crust, where it may account for as much as 5 of its 7-kilometer total thickness.

In this manner, the processes at work along the ridge system generate the entire sequence of rocks found in an ophiolite complex. Since the magma chambers are periodically replenished with fresh magma rising from the asthenosphere, the oceanic crust is continuously being generated.

Interactions between Seawater and Oceanic Crust

In addition to serving as a mechanism for the dissipation of Earth's internal heat, the interaction between seawater and the newly formed basaltic crust alters both the seawater and the crust. Because submarine lava flows are very permeable and the upper oceanic crust is highly fractured, seawater can penetrate to a depth of 2 to 3 kilometers. As seawater circulates through the hot crust, it is heated and alters the basaltic rock by a process called *hydrothermal* (hot water) *metamorphism* (see Chapter 8). This alteration causes the dark silicates (olivine and pyroxene) found in basalt to form new minerals such as chlorite and serpentine.

In addition to the basaltic crust being altered, so is the seawater. As the hot seawater circulates through the newly formed rock, it dissolves ions of silica, iron, copper, and sometimes silver and gold from the hot basalts. Once the water is heated to several hundred degrees Celsius, it buoyantly rises along fractures and eventually spews out at the surface (see Box 13.3). Studies conducted by submersibles along the Juan de Fuca Ridge have photographed these metallic-rich solutions as they gush from the seafloor to form particle-filled clouds called **black smokers.** As the

BOX 13.3 ▸ EARTH AS A SYSTEM

Deep-Sea Hydrothermal Vents*

Sitting in a computer-filled, darkened room, a group of geologists, biologists, and chemists peer intently at video monitors showing astonishing imagery of giant, smoke-billowing, chimney-like rock formations and an abundance of bizarre animals. Is this a scene from Hollywood's latest sci-fi blockbuster? No, it's a typical scene from an actual research vessel located 250 kilometers southwest of Vancouver Island. The images are being relayed to the ship by the Canadian remotely controlled vehicle *ROPOS (Remotely Operated Platform for Ocean Science)*, which is busy working more than 2 kilometers below the surface along the Juan de Fuca Ridge. This seafloor mountain chain is actively being created by the rifting and pulling apart of the Pacific and Juan de Fuca tectonic plates and the production of new oceanic crust by upwelling magma.

Along ridges like the Juan de Fuca, cold seawater moves several hundreds of meters into the highly fractured basaltic crust, where it is then heated at depth by magmatic sources. Along the way, the heated water strips metals and elements such as sulfur from the surrounding rocks. This heated fluid tends to rise due to convection, following conduits and fractures that are concentrated along the ridges. When it reaches the surface of the crust, the fluid can be over 400°C, but it does not boil because of the extremely high pressures caused by the water column above the vents. When this hydrothermal fluid comes into contact with the much colder seawater, minerals rapidly precipitate and form a shimmering smoke-like cloud, which gives "*black smokers*" their name (Figure 13.D). Some minerals immediately solidify and contribute to the formation of spectacular chimney-like structures, which can be as tall as a 15-story building, and are appropriately given names like *Godzilla* and *Inferno*. In some cases, these chimneys and related deposits contain concentrated amounts of iron, copper, zinc, lead, silver, and even gold.

The Juan de Fuca vents are also remarkable for the biology that they support. In these environments completely devoid of sunlight, microorganisms take advantage of the mineral-rich hydrothermal fluid to perform what is known as chemosynthesis. The microbial communities, in turn, support larger, more complex animals such as fish, crabs, worms, mussels, and clams. Some species at these vents are not found anywhere else on Earth. The most famous of these, and perhaps the most unique, is the tubeworm (Figure 13.E). With their white chitinous tubes and bright red plumes, these conspicuous creatures rely entirely on bacteria growing in their *trophosome*, an internal organ designed for harvesting bacteria. These *symbiotic* bacteria rely on the tubeworm to provide them with a suitable habitat and, in return, they provide carbon-based building blocks to the tubeworms.

FIGURE 13.E Tube worms up to 3 meters (10 feet) in length are among the organisms found in the extreme environment of hydrothermal vents along the crest of the oceanic ridge, where sunlight is nonexistent. These organisms obtain their food from internal microscopic bacteria-like organisms, which acquire their nourishment and energy through the processes of chemosynthesis. (Photo by Al Giddings Images, Inc.)

Concern over the possibility of damaging these unique ecosystems by sampling activities of scientists, increasing ecotourism, or the potential exploitation of biological and mineral resources, has recently led the Canadian government to label part of the Juan de Fuca hydrothermal vents as Canada's first Marine Protected Area (MPA). This designation puts in place enforceable regulations to preserve and protect the area and its marine organisms, while encouraging continued scientific study of this unique and remarkable ecosystem located in Canada's backyard.

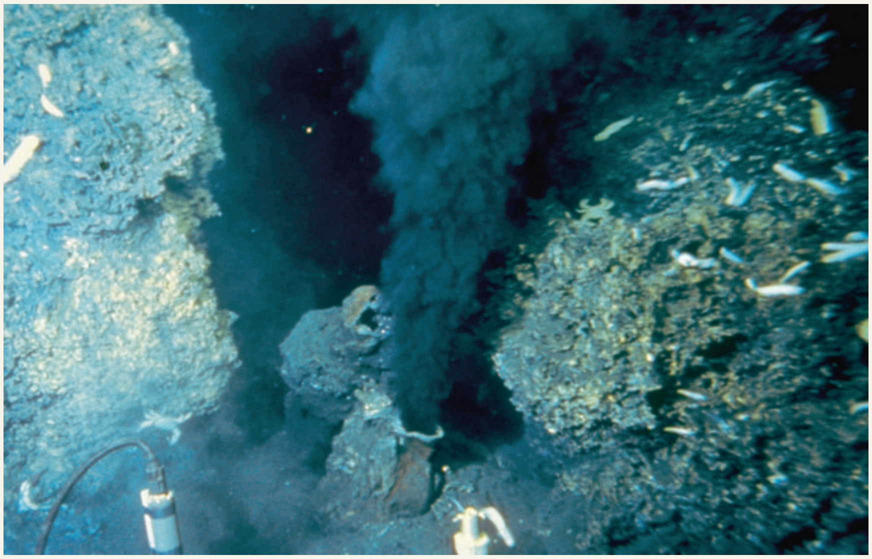


FIGURE 13.D A black smoker spewing hot, mineral-rich water along the East Pacific Rise. As heated solutions meet cold seawater, sulfides of copper, iron, and zinc precipitate immediately, forming mounds of minerals around these vents. (Photo by Dudley Foster, Woods Hole Oceanographic Institution)

*This box was prepared by Richard Leveille, a research scientist at the University of Quebec, Montreal, Canada.

hot liquid (about 350°C) mixes with the cold, mineral-laden seawater, the dissolved minerals precipitate to form massive, metallic sulfide deposits, some of which are economically important. Occasionally these deposits grow upward to form large, chimneylike structures as tall as skyscrapers.

Continental Rifting: The Birth of a New Ocean Basin

GEODe EARTH

Divergent Boundaries

▸ The Formation of Ocean Basins

Why the supercontinent of Pangaea began to split apart nearly 200 million years ago is not known with certainty. Nevertheless, this event serves to illustrate that perhaps most ocean basins get their start when a continent begins to break apart. This clearly is the case for the Atlantic Ocean, which formed as the Americas drifted from Europe and Africa. It is also true for the Indian Ocean, which developed as Africa rifted from Antarctica and India.

FIGURE 13.20 East African rift valleys and associated features.

Evolution of an Ocean Basin

The development of a new ocean basin begins with the formation of a **continental rift,** an elongated depression in which the entire thickness of the lithosphere has been deformed. Examples of continental rifts include the East African Rift, the Baikal Rift (south central Siberia), the Rhine Valley (northwestern Europe), the Rio Grande Rift, and the Basin and Range province in the western United States. It appears that continental rifts form in a variety of tectonic settings and may result in the breakup of a landmass.

In those settings where rifting continues, the rift system will evolve into a young, narrow ocean basin, exemplified by the present-day Red Sea. Eventually, seafloor spreading results in the formation of a mature ocean basin bordered by rifted continental margins. The Atlantic Ocean is such a feature. What follows is a look at this model of ocean basin evolution using modern examples to represent the various stages of rifting.

East African Rift An example of an active continental rift is the East African Rift, which extends through eastern Africa for approximately 3000 kilometers (2000 miles). Rather than being a single rift, the East African Rift consists of several somewhat interconnected rift valleys that split into an eastern and western section around Lake Victoria (Figure 13.20). Whether this rift will develop into a spreading center, where the Somali subplate separates from the continent of Africa, is still being debated. Nevertheless, the East African Rift is thought to characterize the initial stage in the breakup of a continent.

The most recent period of rifting began about 20 million years ago as upwelling in the mantle forcefully intruded the base of the lithosphere (Figure 13.21A). Buoyant uplifting of the heated lithosphere led to doming of the crust. As a consequence, the upper crust was broken along steep-angle normal faults, producing downfaulted blocks, or *grabens,* while the lower crust deformed by ductile stretching (Figure 13.21B). Thus, this continental rift system closely resembles rifts found along slow spreading centers.

In its early stage of formation, magma generated by decompression melting of the rising mantle plume intrudes the crust. Some of the magma migrates along fractures and erupts at the surface. This activity produces extensive basaltic flows within the rift as well as volcanic cones—some forming more than 100 kilometers from the rift axis. Examples include Mount Kilimanjaro, which is the highest point in Africa, rising almost 6000 meters (20,000 feet) above the Serengeti Plain, and Mount Kenya.

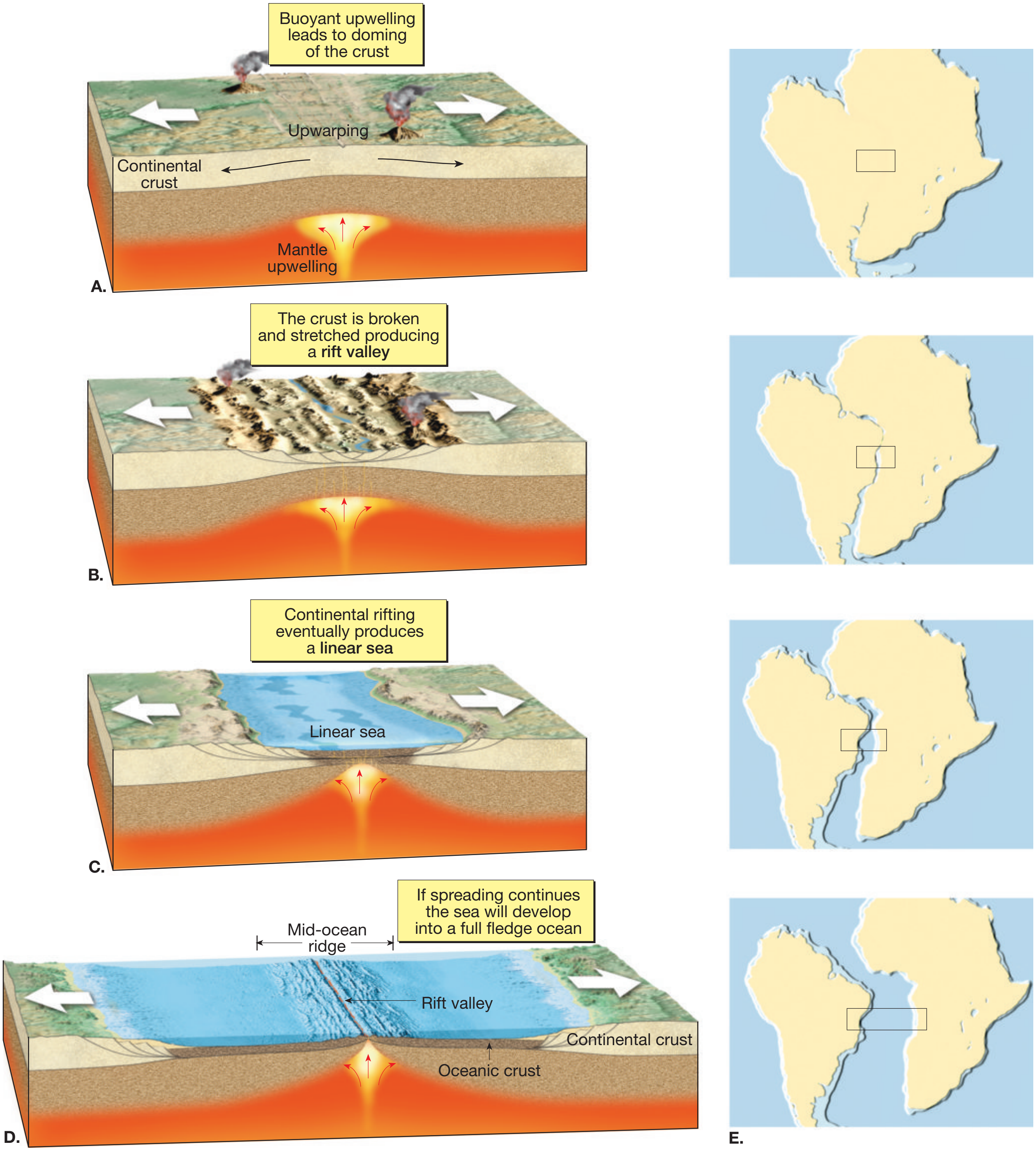


FIGURE 13.21 Formation of an ocean basin. **A.** Tensional forces and buoyant uplifting of the heated lithosphere cause the upper crust to be broken along normal faults, while the lower crust deforms by ductile stretching. **B.** As the crust is pulled apart, large slabs of rock sink, generating a rift zone. **C.** Further spreading generates a narrow sea. **D.** Eventually, an expansive ocean basin and ridge system are created. **E.** Illustration of the separation of South America and Africa to form the South Atlantic.

Red Sea Research suggests that if tensional forces are maintained, a rift valley will lengthen and deepen, eventually extending out to the margin of the continent, thereby splitting it in two (Figure 13.21C). At this point, the continental rift becomes a narrow linear sea with an outlet to the ocean, similar to the Red Sea.

The Red Sea formed when the Arabian Peninsula rifted from Africa beginning about 30 million years ago. Steep fault scarps that rise as much as 3 kilometers above sea level flank the margins of this water body. Thus, the escarpments surrounding the Red Sea are similar to the cliffs that border the East African Rift. Although the Red Sea only reaches oceanic depths (up to 5 kilometers) in a few locations, symmetrical magnetic stripes indicate that typical seafloor spreading here has been taking place for the past 5 million years.

Atlantic Ocean If spreading continues, the Red Sea will grow wider and develop an elevated oceanic ridge similar to the Mid-Atlantic Ridge (Figure 13.21D). As new oceanic crust is added to the diverging plates, the rifted continental margins move ever so slowly away from one another. As a result, the rifted continental margins that were once situated above the region of upwelling are displaced toward the interior of the growing plates. Consequently, as the continental lithosphere moves away from the source of heat, it cools, contracts, and subsides.

In time these continental margins will subside below sea level. Simultaneously, material eroded from the adjacent landmass will be deposited atop the faulted topography of the submerged continental margin. Eventually, this material will accumulate to form a thick, broad wedge of relatively undisturbed sediment and sedimentary rock. Recall that continental margins of this type are called *passive continental margins.* Because passive margins are not associated with plate boundaries, they experience little volcanism and few earthquakes. Recall, however, that this was not the case when these lithospheric blocks made up the flanks of a continental rift.

Not all continental rift valleys develop into full-fledged spreading centers. Running through the central United States is a failed rift that extends from Lake Superior into central Kansas (Figure 13.22). This once active rift valley is filled with volcanic rock that was extruded onto the crust more than a billion years ago. Why one rift valley develops into an active spreading center while others are abandoned is not yet known.

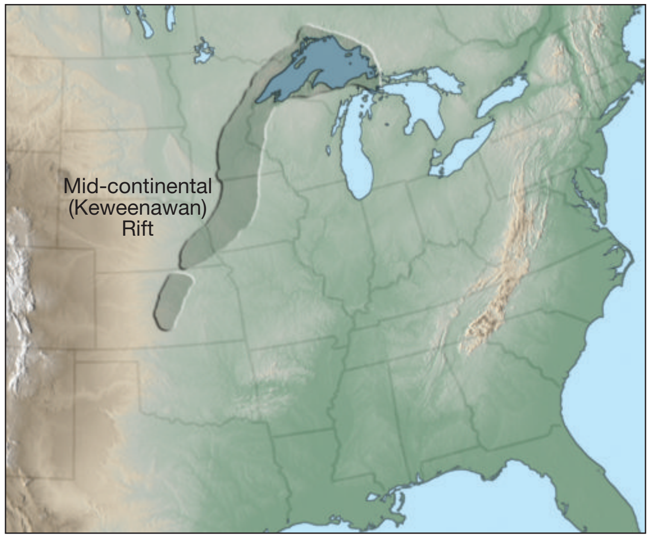


FIGURE 13.22 Map showing location of a failed rift extending from Lake Superior to Kansas.

Mechanisms for Continental Rifting

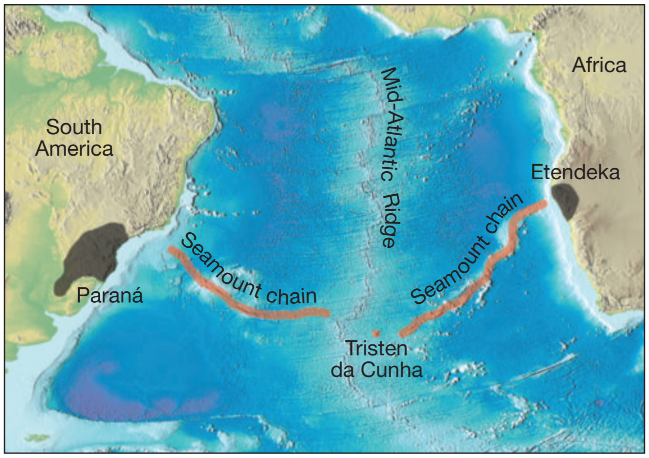
It seems likely that supercontinents existed sporadically during the geologic past. Pangaea, which was the most recent of these, was assembled into a supercontinent between 450 and 230 million years ago, only to break up again shortly thereafter. Thus, geologists have concluded that the formation of a supercontinent followed by continental splitting must be an integral part of plate tectonics. Furthermore, this phenomenon must involve a major change in the direction and nature of the forces that drive plate motion. Stated another way, over long periods of geologic time, the forces that drive plate motions tend to organize crustal fragments into a single supercontinent, only to change directions and disperse them again. Two mechanisms have been proposed for continental rifting—plumes of hot mobile rock rising from deep in the mantle, and forces that arise from plate motions.

Mantle Plumes and Gravity Sliding Recall that a *mantle plume* consists of hotter than normal mantle rock that has a large mushroom-shaped head hundreds of kilometers in diameter attached to a long, but narrow, trailing tail. As the plume head nears the base of the cool lithosphere, it spreads laterally. Decompression melting within the plume head generates huge volumes of basaltic magma that rises and triggers volcanism at the surface. The result is a volcanic region, called a *hot spot,* that can be as much as 2000 kilometers across.

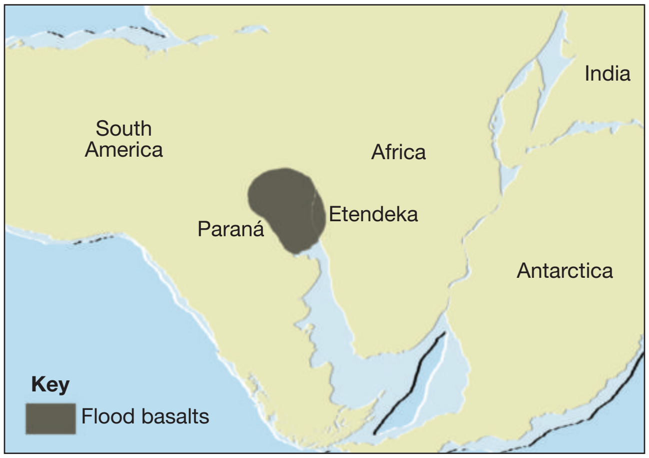
Research suggests that mantle plumes would tend to concentrate beneath a supercontinent, because once assembled, a large landmass forms an insulating "blanket" that traps heat in the mantle. The resulting temperature increase would lead to the formation of mantle plumes that serve as heat dissipation mechanisms.

Evidence for the role that mantle plumes play in continental rifting is available from passive continental margins, the former sites of rifting. In several regions on both sides of the Atlantic, continental rifting was preceded by crustal uplift and massive outpourings of basaltic lava. Examples include the Etendeka flood basalts of southwest Africa and the Paraná basalt province of South America (Figure 13.23A).

About 130 million years ago, when South America and Africa were joined as a single landmass, vast outpourings of lava produced a large continental basalt plateau (Figure 13.23B). Shortly after this event, the South Atlantic began to open, splitting the basalt province into what is now the Etendeka and Paraná basalt plateaus. As the ocean basin grew, the tail of the plume produced a string of seamounts



A.



B.

FIGURE 13.23 Evidence for the role that mantle plumes might play in continental rifting. **A.** Relationship of the Paraná and Etendeka basalt plateaus to the Tristan da Cunha hotspot. **B.** Location of these basalt plateaus 130 million years ago, just before the South Atlantic began to open.

on each side of the newly formed ridge (Figure 13.23A). The modern area of hot-spot activity is centered around the volcanic island of Tristan da Cunha, which is located on the Mid-Atlantic Ridge.

About 60 million years ago, yet another mantle plume is thought to have initiated the rifting of Greenland from Northern Europe. Volcanic rocks associated with this activity extend from eastern Greenland to Scotland. The hot spot associated with this event is presently situated beneath Iceland.

From these studies, geologists have concluded that mantle plumes have played a role in the development of at least some continental rifts. In these regions, rifting began when a hot mantle plume reached the base of the lithosphere and caused the overlying crust to dome and weaken. As the crust was buoyantly uplifted, it stretched and developed rifts similar to those in East Africa. Simultaneously, decompression melting of the plume head led to vast outpourings of basaltic lavas. Following these episodes of igneous activity, an ocean basin began to open. The proposed mechanism for rifting is gravity sliding off the uplift caused by plume buoyancy.

It is important to note that not all hot-spot volcanism leads to rifting. For example, vast outpourings of basaltic lava that constitute the Columbia River basalts in the Pacific Northwest, as well as Russia's Siberian Traps, are not associated with the fragmentation of a continent.

Slab Pull and Slab Suction It is generally agreed that tensional forces, which tend to elongate or pull apart a rock unit, are needed in order for a continent to be fragmented. But how do these forces originate?

Recall that old oceanic lithosphere subducts because it is denser than the underlying asthenosphere. In situations where a continent is attached to a subducting slab of oceanic lithosphere, it will be pulled toward the trench. However, continents overlie thick sections of lithospheric mantle. As a result, they tend to resist being towed, which creates tensional stresses that stretch and thin the crust. Whether slab pull can tear a continent apart is still being studied. Perhaps other factors, including the presence of hot spots, or an inherent weakness in the crust, such as a major fault zone, may contribute to rifting.

Investigators have suggested that during the breakup of Pangaea, the Americas were rifted from Europe and Africa as a result of another force—*slab suction.* Recall that when a cold oceanic slab sinks, it causes the trench to retreat, or roll back. This creates flow in the asthenosphere that pulls the overriding plate *toward* the retreating trench (Figure 13.24).

During the breakup of Pangaea a subduction zone extended along the entire western margin of North and South America. As this subduction zone developed, the trench slowly retreated westward toward a spreading center located in the Pacific. Modern remnants of this subduction zone include the Peru–Chile Trench, Central American Trench, and Cascadia subduction zone (see Figure 13.9, p. 357). Slab suction along the entire western margin of the Americas may have provided the tensional forces that rifted Pangaea.

In summary, continental rifting occurs when a landmass is under tension, which tends to elongate and thin the lithosphere. This mechanism may be aided by a series of hotspots that weaken and elevate the crust.

Destruction of Oceanic Lithosphere

Although new lithosphere is continually being produced at divergent plate boundaries, Earth's surface area is not growing larger. In order to balance the amount of newly created lithosphere, there must be a process whereby plates are destroyed. Recall that this occurs along *convergent boundaries,* also called *subduction zones.*

Why Oceanic Lithosphere Subducts

The process of plate subduction is complex, and the ultimate fate of oceanic lithosphere is still being debated. What is known with some certainty is that a slab of oceanic lithosphere subducts because its overall density is greater than that of the underlying mantle. Recall that when ocean

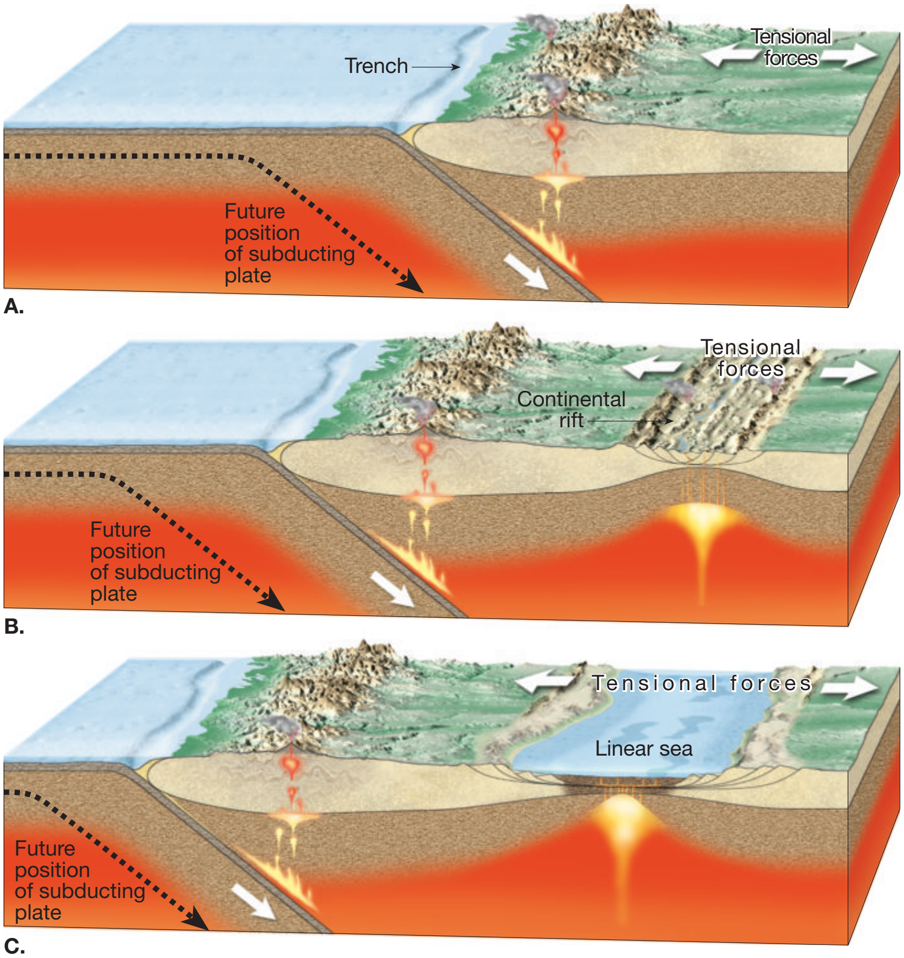


FIGURE 13.24 Illustration of how trench retreat, or roll back, produces slab-suction forces that are thought to contribute to the breakup of a continent.

It is important to note that it is the *lithospheric mantle,* located beneath the oceanic crust, that drives subduction. Even when the oceanic crust is quite old, its density is 3.0 g/cm^3, which is less than the underlying asthenosphere with a density of about 3.2 g/cm^3. Only because the cold lithospheric mantle is denser than the warmer asthenosphere that supports it does subduction occur.

At some locations, the oceanic crust is unusually thick because it contains a chain of seamounts. Here the lithosphere may have sufficient crustal material, and hence buoyancy, to prevent or at least modify subduction. This appears to be the situation in two areas along the Peru–Chile Trench, where the angle of descent is quite shallow—about 10 to 15 degrees. Low dip angles often result in a strong interaction between the descending slab and the overriding plate. Consequently, these regions experience frequent, great earthquakes.

It has also been determined that unusually thick units of oceanic crust, those that are greater than 30 kilometers in thickness, probably will not subduct. An example is the Ontong Java Plateau, which is a thick oceanic basalt plateau located in the western Pacific. About 20 million years ago this plateau reached the trench that formed the boundary between the subducting Pacific plate and the overriding Australian–Indian plate. Apparently too buoyant to subduct, the Ontong Java Plateau clogged the trench and shut down subduction at this site. We will consider what eventually happens to these crustal fragments that are too buoyant to subduct in the next chapter.

crust forms along a ridge, it is warm and buoyant, a fact that results in the ridge being elevated above the deep-ocean basins. However, as oceanic lithosphere moves away from the site of warm upwelling, it cools and thickens. After about 15 million years, an oceanic slab tends to be denser than the supporting asthenosphere. In parts of the western Pacific some oceanic lithosphere is nearly 180 million years old. This is the thickest and most dense in today's oceans. The subducting slabs in this region typically descend into the mantle at angles approaching 90 degrees (Figure 13.25A). Sites where plates subduct at such steep angles are found in association with the Tonga, Mariana, and Kurile trenches.

When a spreading center is located near a subduction zone, the oceanic lithosphere is still young and therefore warm and buoyant. Hence, the angle of descent for these slabs is small (Figure 13.25B). It is even possible that oceanic lithosphere may be overridden by a continental landmass before it has cooled sufficiently to readily subduct. In this situation, the slab may be so buoyant that rather than plunging into the mantle, it moves horizontally beneath a block of continental lithosphere. This phenomenon is called **buoyant subduction.** Buoyant slabs are thought to eventually sink when they cool sufficiently and their density increases.

Subducting Plates: The Demise of an Ocean Basin

Using magnetic stripes and fracture zones on the ocean floor, geologists began reconstructing the movement of plates over the past 200 million years. From this work they discovered that parts, or even entire ocean basins, have been destroyed along subduction zones. For example, during the breakup of Pangaea shown in Figure 2.A (p. 42), notice that the African plate rotates and moves northward. Eventually the northern margin of Africa collides with Eurasia. During this event, the floor of the intervening Tethys Ocean was almost entirely consumed into the mantle, leaving behind only a small remnant—the Mediterranean Sea.

Reconstructions of the breakup of Pangaea also helped investigators understand the demise of the Farallon plate—a large oceanic plate that once occupied much of the eastern Pacific basin. Prior to the breakup, the Farallon plate, plus one or two smaller plates, were situated opposite the Pacific plate on the eastern side of a spreading center located near

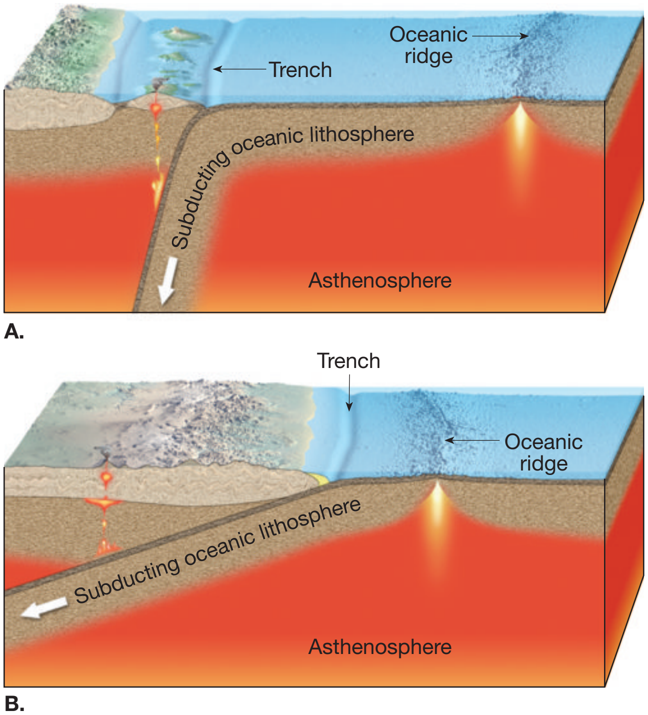


FIGURE 13.25 The angle at which oceanic lithosphere descends into the asthenosphere depends on its density. **A.** In parts of the Pacific, some oceanic lithosphere is older than 160 million years and typically descends into the mantle at angles approaching 90 degrees. **B.** Young oceanic lithosphere is warm and buoyant, hence it tends to subduct at a low angle.

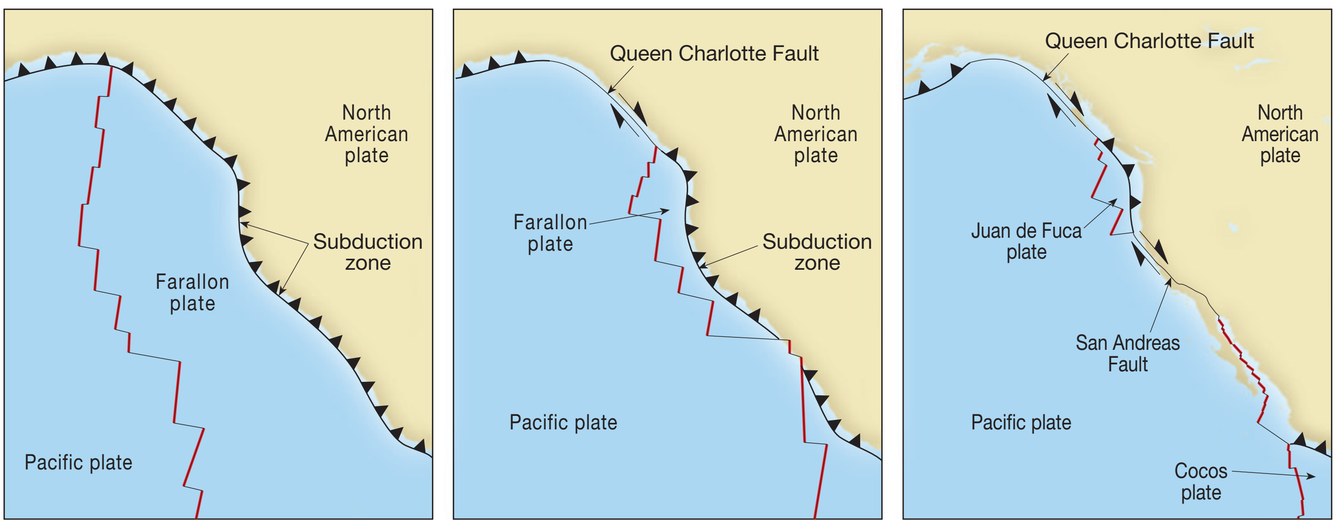
the center of the Pacific basin. A modern remnant of this spreading center, which generated both the Farallon and Pacific plates, is the East Pacific Rise.

Beginning about 180 million years ago, the Americans were propelled westward by seafloor spreading in the Atlantic. Hence, the convergent plate boundaries that formed along the west coasts of North and South America gradually migrated westward relative to the spreading center located in the Pacific. The Farallon plate, which was subducting beneath the Americans faster than it was being generated, got smaller and smaller (Figure 13.26). As its surface area decreased, it broke into smaller pieces, some of which subducted entirely. The remaining fragments of the once mighty Farallon plate are the Juan de Fuca, Cocos, and Nazca plates.

As the Farallon plate shrank, the Pacific plate grew larger, encroaching on the American plates. About 30 million years ago, a section of the East Pacific Rise collided with the subduction zone that once lay off the coast of California (Figure 13.26B). As this spreading center subducted into the California trench, these structures were mutually destroyed and replaced by a newly generated transform fault system that accommodates the differential motion between the North American and Pacific plates. As more of the ridge was subducted, the transform fault system, which we now call the San Andreas Fault, propagated through western California (Figure 13.26). Farther north, a similar event generated the Queen Charlotte transform fault.

Consequently, much of the present boundary between the Pacific and North American plates lies along transform faults located within the continent. In the United States (outside of Alaska), the only remaining part of the extensive convergent boundary that once ran along the entire West Coast is the Cascadia subduction zone. Here the subduction of the Juan de Fuca plate has generated the volcanoes of the Cascade Range.

Today, the southern end of the San Andreas Fault connects to a young spreading center (an extension of the East Pacific Rise) that generated the Gulf of California (Figure 13.27). Because of this change in plate geometry, the Pacific plate has captured a sliver of North America (the Baja Peninsula) and is carrying it northwestward toward Alaska at a rate of about 6 centimeters per year.



A. 56 million years ago **B. 37 million years ago** **C. Today**

FIGURE 13.26 Simplified illustration of the demise of the Farallon plate, which once ran along the western margin of the Americas. Because the Farallon plate was subducting faster than it was being generated, it got smaller and smaller. The remaining fragments of the once mighty Farallon plate are the Juan de Fuca, Cocos, and Nazca plates.

FIGURE 13.27 Satellite image showing the separation between the Baja Penninsula and North America. (Image courtesy of NASA)

Students Sometimes Ask . . .

In class you stated that when oceanic lithosphere reaches about 15 million years in age, it will have cooled sufficiently to be denser than the underlying asthenosphere. Why doesn't it begin to subduct at that point?

Plate-mantle convection is much more complicated than classic convective flow that develops when a liquid is heated from below. In a convective liquid, as soon as the material at the top cools and becomes more dense than the material below, it begins to sink. In plate-mantle convection, the upper boundary layer—the lithosphere—is a rigid solid. For a new subduction zone to develop, an area of weakness needs to exist somewhere in the lithospheric slab. In addition, the negative buoyancy of the lithosphere must be sufficient to overcome the strength of the cool rigid plate. Stated another way, in order for part of a plate to be subducted, the forces acting on the plate must be large enough to bend the plate.

Summary

- *Ocean bathymetry is determined using echo sounders and multibeam sonars,* which bounce sonic signals off the ocean floor. Ship-based receivers record the reflected echoes and accurately measure the time interval of the signals. With this information, ocean depths are calculated and plotted to produce maps of ocean-floor topography. Recently, *satellite measurements* of the ocean surface have added data for mapping ocean-floor features.
- Oceanographers studying the topography of the ocean basins have delineated three major units: *continental margins, deep-ocean basins*, and *oceanic (mid-ocean) ridges.*
- The zones that collectively make up a *passive continental margin* include the *continental shelf* (a gently sloping, submerged surface extending from the shoreline toward the deep-ocean basin); *continental slope* (the true edge of the continent, which has a steep slope that leads from the continental shelf into deep water); and in regions where trenches do not exist, the relatively steep continental slope merges into a more gradual incline known as the *continental rise.* The continental rise consists of sediments that have moved downslope from the continental shelf to the deep-ocean floor.
- *Active continental margins* are located primarily around the Pacific Ocean in areas where the leading edge of a continent is overrunning oceanic lithosphere. Here sediment scraped from the descending oceanic plate is plastered against the continent to form a collection of sediments called an *accretionary wedge.* An active continental margin generally has a narrow continental shelf, which grades into a deep-ocean trench.
- The deep-ocean basin lies between the continental margin and the oceanic ridge system. Its features include *deep-ocean trenches* (long, narrow depressions that are the deepest parts of the ocean and are located where moving crustal plates descend back into the mantle); *abyssal plains* (among the most level places on Earth, consisting of thick accumulations of sediments that were deposited atop the low, rough portions of the ocean floor by turbidity currents); *seamounts* (volcanic peaks on the ocean floor that originate near oceanic ridges or in association with volcanic hot spots); and *oceanic plateaus* (large flood basalt provinces similar to those found on the continents).
- *Oceanic (mid-ocean) ridges,* the sites of seafloor spreading, are found in all major oceans and represent more than 20 percent of Earth's surface. They are the most prominent features in the oceans, because they form an almost continuous swell that rises 2 to 3 kilometers above the adjacent ocean basin floor. Ridges are characterized by an *elevated position*, *extensive faulting,* and *volcanic structures*

that have developed on newly formed oceanic crust. Most of the geologic activity associated with ridges occurs along a narrow region on the ridge crest, called the *rift zone*, where magma from the asthenosphere moves upward to create new slivers of oceanic crust. The topography of the various segments of the oceanic ridge is controlled by the rate of seafloor spreading.

- New oceanic crust is formed in a continuous manner by the process of seafloor spreading. The upper crust is composed of *pillow lavas* of basaltic composition. Below this layer are numerous interconnected dikes (*sheeted dike complex*) that are underlain by a thick layer of gabbro. This entire sequence is called an *ophiolite complex*.
- The development of a new ocean basin begins with the formation of a *continental rift* similar to the East African Rift. In those settings where rifting continues, a young, narrow ocean basin develops, exemplified by the Red Sea. Eventually, seafloor spreading creates an ocean basin bordered by rifted continental margins similar to the present-day Atlantic Ocean. Two mechanisms have been proposed for continental rifting—plumes of hot mobile rock rising from deep in the mantle and forces that arise from plate motions.
- Oceanic lithosphere subducts because its overall density is greater than the underlying asthenosphere. The subduction of oceanic lithosphere may result in the destruction of parts—or even entire—ocean basins. A classic example is the Farallon plate, most of subducted beneath the American plates as they were displaced westward by seafloor spreading in the Atlantic.

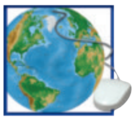
Review Questions

1. Assuming that the average speed of sound waves in water is 1500 meters per second, determine the water depth if the signal sent out by an echo sounder requires 6 seconds to strike bottom and return to the recorder (see Figure 13.2).
2. Describe how satellites orbiting Earth can determine features on the seafloor without being able to directly observe them beneath several kilometers of seawater.
3. What are the three major topographic provinces of the ocean floor?
4. List the three major features that comprise a passive continental margin. Which of these features is considered a flooded extension of the continent? Which one has the steepest slope?
5. Describe the differences between active and passive continental margins. Be sure to include how various features relate to plate tectonics, and give a geographic example of each type of margin.
6. Why are abyssal plains more extensive on the floor of the Atlantic than on the floor of the Pacific?
7. How does a flat-topped *seamount*, or *guyot*, form?
8. Briefly describe the oceanic ridge system.
9. Although oceanic ridges can be as tall as some mountains found on the continents, how are these features different?
10. What is the source of magma for seafloor spreading?
11. What is the primary reason for the elevated position of the oceanic ridge system?
12. How does hydrothermal metamorphism alter the basaltic rocks that make up the seafloor? How is seawater changed during this process?
13. What is a black smoker?
14. Compare and contrast a slow spreading center such as the Mid-Atlantic Ridge with one that exhibits a faster spreading rate, such as the East Pacific Rise.
15. Briefly describe the four layers of the ocean crust.
16. How does the *sheeted dike complex* form? What about the lower unit?
17. Name a place that exemplifies a continental rift.
18. What role are mantle plumes thought to play in the rifting of a continent?
19. What evidence suggests that hotspot volcanism does not always lead to the breakup of a continent?
20. Explain why oceanic lithosphere subducts even though the oceanic crust is less dense than the underlying asthenosphere.
21. Why does the lithosphere thicken as it moves away from the ridge as a result of seafloor spreading?
22. What happened to the Farallon plate? Name the remaining parts.

Key Terms

abyssal plain (p. 357)
accretionary wedge (p. 356)
active continental margin (p. 355)
bathymetry (p. 350)
black smokers (p. 364)
buoyant subduction (p. 370)
continental margin (p. 353)
continental rift (p. 366)
continental rise (p. 355)
continental shelf (p. 353)
continental slope (p. 355)
deep-ocean basin (p. 356)
deep-ocean trench (p. 356)
deep-sea fan (p. 355)
echo sounder (p. 350)
guyot (p. 358)
mid-ocean ridge (p. 358)
oceanic plateau (p. 358)
oceanic ridge (p. 358)
ophiolite complex (p. 363)
passive continental margin (p. 353)
pillow basalts (p. 364)
rift valley (p. 361)
seamount (p. 357)
seismic reflection profile (p. 352)
sheeted dike complex (p. 364)
sonar (p. 350)
tablemount (p. 358)

Web Resources



The *Earth* Website uses the resources and flexibility of the Internet to aid in your study of the topics in this chapter. Written and developed by geology instructors, this site will help improve your understanding of geology. Visit **http://www.prenhall.com/tarbuck** and click on the cover of *Earth 9e* to find:

- Online review quizzes.
- Critical thinking exercises.
- Links to chapter-specific Web resources.
- Internet-wide key-term searches.

http://www. prenhall.com/tarbuck

GEODe: Earth

GEODe: Earth makes studying faster and more effective by reinforcing key concepts using animation, video, narration, interactive exercises and practice quizzes. A copy is included with every copy of *Earth.*