CHAPTER 4

Plate Tectonics: Evolution of the Ocean Floor

CRITICAL CONCEPTS USED IN THIS CHAPTER

CC1 Density and Layering in Fluids
CC2 Isostasy, Eustasy, and Sea Level
CC3 Convection and Convection Cells
CC7 Radioactivity and Age Dating
CC11 Chaos
CC14 Phototrophy, Light, and Nutrients

FIGURE 4-1. Pu'u'ō'ō, the active vent of the Kilauea Volcano in Hawaii, is the result of a hot spot. This volcano often erupts relatively smoothly, but in 1997 it ejected cinders and ash into the atmosphere that "rained" on the airplane windshield like a hailstorm. The shiny surface below the vent mouth is a crust of solidified lava that conceals a rapidly flowing molten river of lava underneath. The river of lava extended several kilometers down the slopes of Kilauea, where it destroyed several homes shortly after this photograph was taken.

The ocean floor has irregular and complex **topographic** features—including mountain ranges, plains, depressions, and plateaus—that resemble topographic features on land (**Fig. 3-4**). During human history, those features have remained essentially unchanged aside from local modifications due to **erosion**, **sedimentation**, volcanic eruptions, and **coral reef** formation. However, the human **species** has existed for only a very brief period of the planet's history (**Fig. 1-4**). In the billions of years before humans appeared, the face of the Earth was reshaped a number of times.

STRUCTURE OF THE EARTH

The continents and ocean basins are not permanent. Instead, the location, shape, and size of these features are continuously changing, albeit imperceptibly slowly on the timescale of human experience. The global-scale processes that continuously reshape the face of the planet are, as yet, far from fully understood. However, we do know that these processes, called **plate tectonics**, originate deep within the Earth.

Layered Structure of the Earth

The Earth consists of a spherical central core surrounded by several concentric layers of different materials (**Fig. 4-2**). The layers are arranged by density, with the highest-density material at the Earth's center and the lowest-density material forming the outer layer, the Earth's crust. This arrangement came about early in the Earth's history, when the planet was much hotter than it is today and almost entirely fluid. The densest elements sank toward the Earth's center and lighter elements rose to the surface (**CC1**). The Earth's center is still much hotter than its surface. Heat is generated continuously within the Earth, primarily by the decay of radioisotopes (**CC7**).

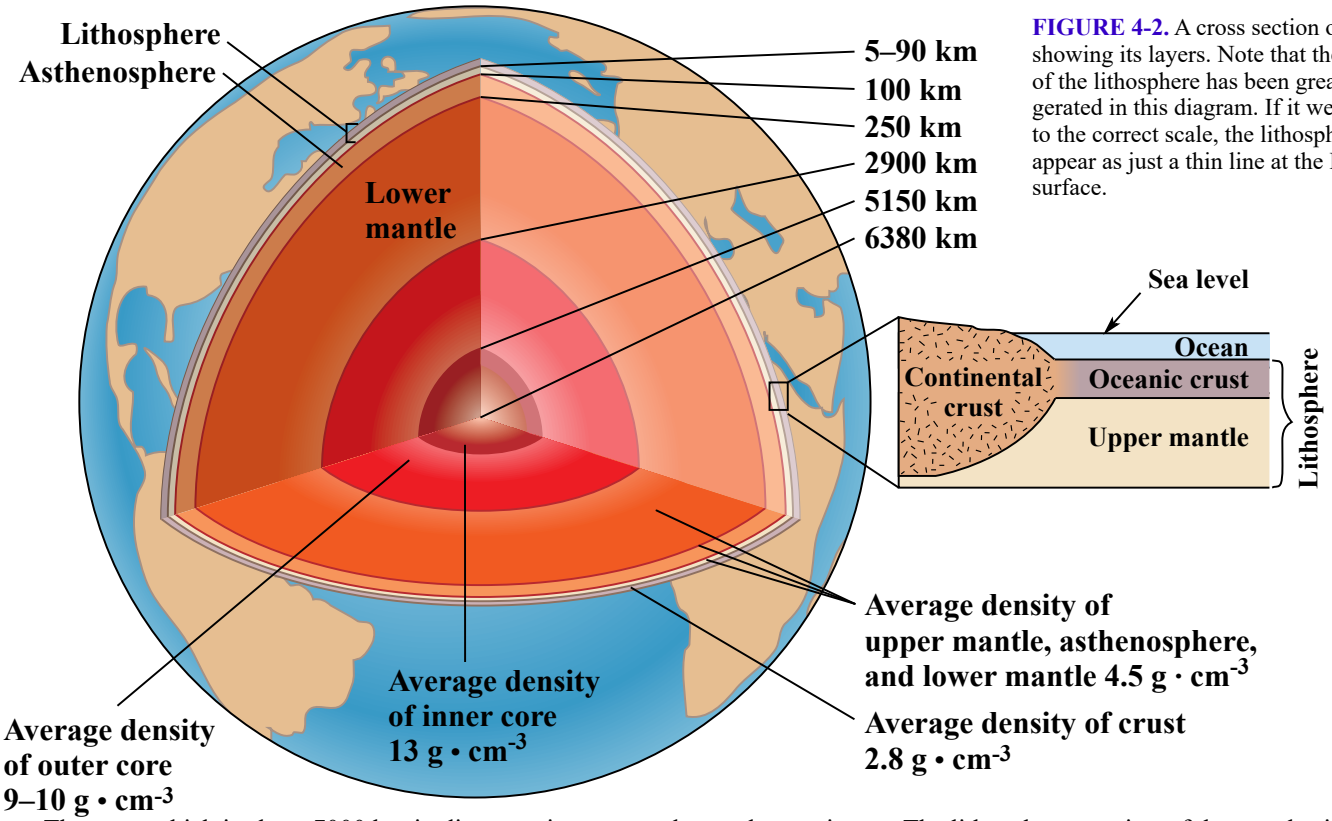

FIGURE 4-2. A cross section of the Earth showing its layers. Note that the thickness of the lithosphere has been greatly exaggerated in this diagram. If it were drawn to the correct scale, the lithosphere would appear as just a thin line at the Earth's surface.

The core, which is about 7000 km in diameter, is composed primarily of iron and nickel, and is very dense. It consists of a solid inner core and a liquid outer core. The **mantle**, which surrounds the core, is composed of material that is about half as dense as the core. Temperature and pressure both increase with depth below Earth's surface. As a result, the upper mantle, known as the **asthenosphere**, is thought to consist of material that is close to its melting point and is "plastic," and capable of flowing very slowly without fracturing. The best example of such materials in common experience may be **glacial** ice. Glaciers flow slowly in response to **gravity**. Within a glacier, pressure due to the weight of the overlying ice changes the properties of the ice deeper than about 50 m so that it too becomes plastic and can flow. Much of the deeper mantle is also thought to be capable of flowing very slowly. The ability to flow is a critically important factor in the tectonic processes that shape the Earth's surface.

The asthenosphere is surrounded by the **lithosphere**, the outermost layer of the Earth, which varies from just a few kilometers in thickness at the **oceanic ridges** to about 100 km in the older parts of the ocean basins and from about 130 to 190 km under the continents. The lithosphere consists of the mostly rigid outer shell of the mantle plus the solid crust that lies on the mantle. The lithosphere is less dense than the asthenosphere and essentially floats on top of the plastic asthenosphere. The oceans and atmosphere lie on top of the lithosphere. Pieces of lithosphere are rigid, but they move across the Earth's surface and in relation to each other as they float on the asthenosphere. They can be many thousands of kilometers wide, but they are generally less than 200 km thick. Because of these "platelike" qualities, they are called **lithospheric plates** (or "tectonic plates" or just "plates"). Processes that occur where the plates collide, move apart, or "slide" past each other are the principal processes that create the mountains, trenches, and other surface features of the continents and the ocean floor.

Lithosphere, Hydrosphere, and Atmosphere

Relative to the Earth, the lithosphere is thin, rather like the skin on an apple (Fig. 4-2). At the top of this thin layer are the mountains, ocean basins, and other features of the Earth's surface. From the deepest point in the ocean to the top of the highest mountain is a vertical distance of approximately 20 km. This 20-

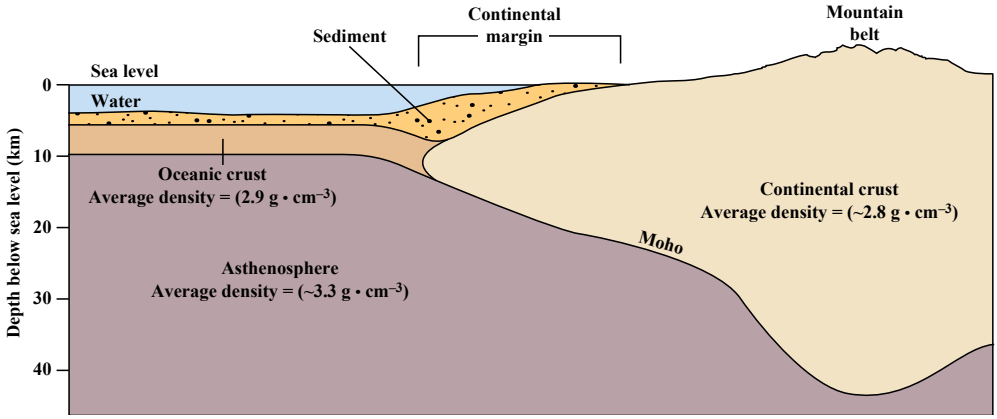

FIGURE 4-3. Structure of the lithosphere. Continental crust is typically 35 to 40 km thick, whereas oceanic crust is typically only 6 to 7 km thick. Continental crust has a lower density than oceanic crust, but both continental and oceanic crust have a considerably lower density than upper mantle material.

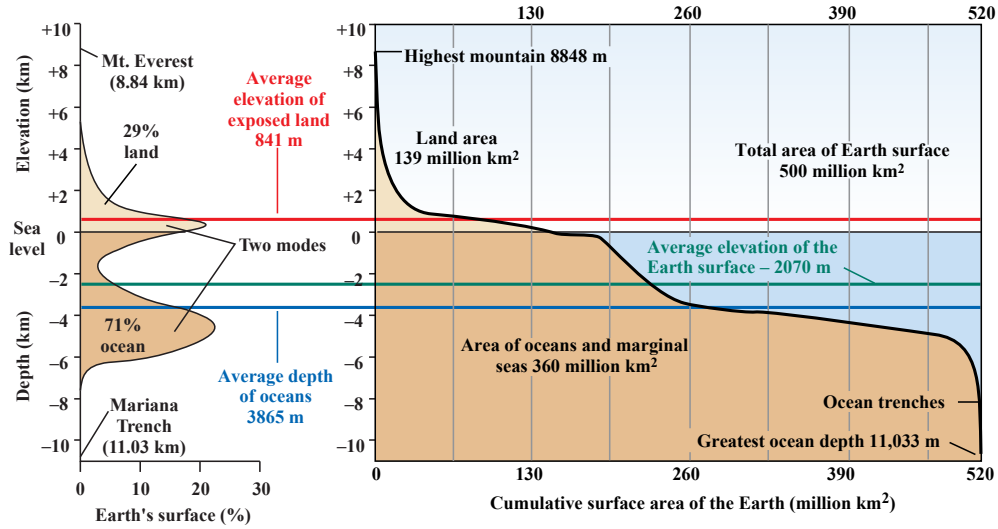

FIGURE 4-4. Relative heights of parts of the Earth's surface. The average elevation of the continents is 841 m above sea level, and the average depth of the oceans is 3865 m. Almost 30% of the Earth's surface is above sea level. Much of the surface of the continental crust is below sea level.

km range is very small compared to the Earth's radius, which is more than 6000 km. Consequently, the planet is an almost smooth sphere (from space it looks smoother than the skin of an orange) on which the mountain ranges and ocean depths are barely perceptible.

There are two types of crust—oceanic and continental—both of which have a substantially lower density than the upper mantle material on which they lie. Oceanic crust has a higher density than continental crust (**Fig. 4-3**). According to the principle of **isostasy** (**CC2**), lithospheric plates float on the asthenosphere at levels determined by their density. Consequently, if the continental and oceanic crusts were of equal thickness, the ocean floor would be lower than the surface of the continents but not as deep as they actually are. However, oceanic crust is much thinner (typically 6–7 km thick) than continental crust (typically 35–40 km), causing an additional height difference between the surface of the continents and the ocean floor (**Fig. 4-4**).

The density difference between continental and oceanic crust is due to differences in their composition. Both are composed primarily of rocks formed from cooling **magma** and consist mainly of silicon and aluminum oxides. However, continental crust is primarily granite, and oceanic crust is primarily gabbro or **basalt**, both of which have higher concentrations of heavier elements, such as iron, and thus a higher density than granite.

Surrounding each continent is a **continental shelf** covered by shallow ocean waters. The continental shelf is an extension of the continent itself, so this portion of the continental crust surface is submerged below sea level. Both the width and the depth of the shelf vary, but it is generally in waters less than 100 to 200 m deep and ranges in width from a few kilometers to several hundred kilometers. The continental shelf slopes gently offshore to the **shelf break**, where it joins the steeper **continental slope** (**Figs. 4-4, 4-5**). The continental slope generally extends to depths of 2 to 3 km. Seaward of the base of the slope, the ocean floor either descends sharply into a deep-ocean **trench** (**Fig. 4-5a**) or slopes gently seaward on a continental rise that eventually joins the deep-ocean floor (**Figs. 4-4, 4-5b**). Much of the deep-ocean floor is featureless flat **abyssal plain**, but other areas are characterized by low, rolling **abyssal hills**.

A layer of **sediment** lies on top of the oceanic crustal rocks, constituting part of the crust. The thickness of the sediment is highly variable, for reasons discussed in this chapter and **Chapter 6**.

The **hydrosphere** consists of all water in the lithosphere that is not combined in rocks and minerals—primarily the oceans and the much smaller volume of freshwater (**Table 5-1**). The oceans cover all the oceanic crust and large areas of continental crust around the edges of the continents; all of which total more than two-thirds of the Earth's surface area.

The atmosphere is the envelope of gases surrounding the lithosphere and hydrosphere and is composed primarily of nitrogen (78%) and oxygen (21%). Although these gases have much lower densities than liquids or solids, they are dense enough to be held by the Earth's gravity. Less dense gases, including hydrogen and helium, are so light that they tend to escape from the Earth's gravity into space. Although the light gases were present in large quantities when the Earth first formed, only trace concentrations are present in the atmosphere today. The atmosphere is discussed in **Chapter 7**.

Studying the Earth's Interior

The processes that occur beneath the lithospheric plates are very difficult to study because they occur deep within the Earth, beneath kilometers of crustal rocks and upper mantle. Studies of the processes below the crust rely primarily on indirect

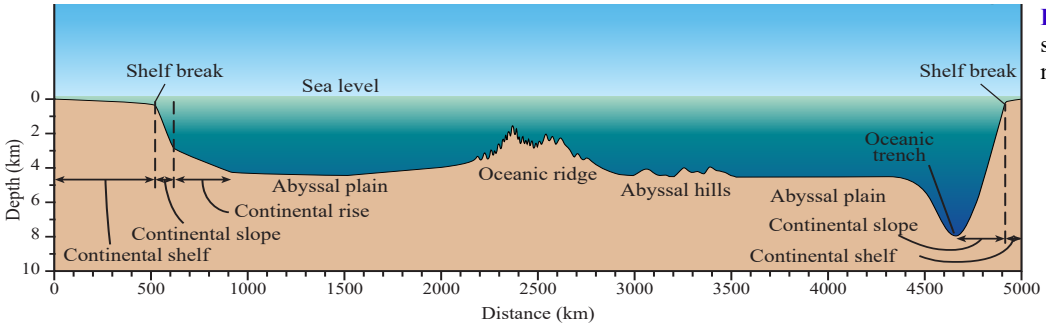

FIGURE 4-5. General features of the seafloor topography between the continents and the deep ocean floor.

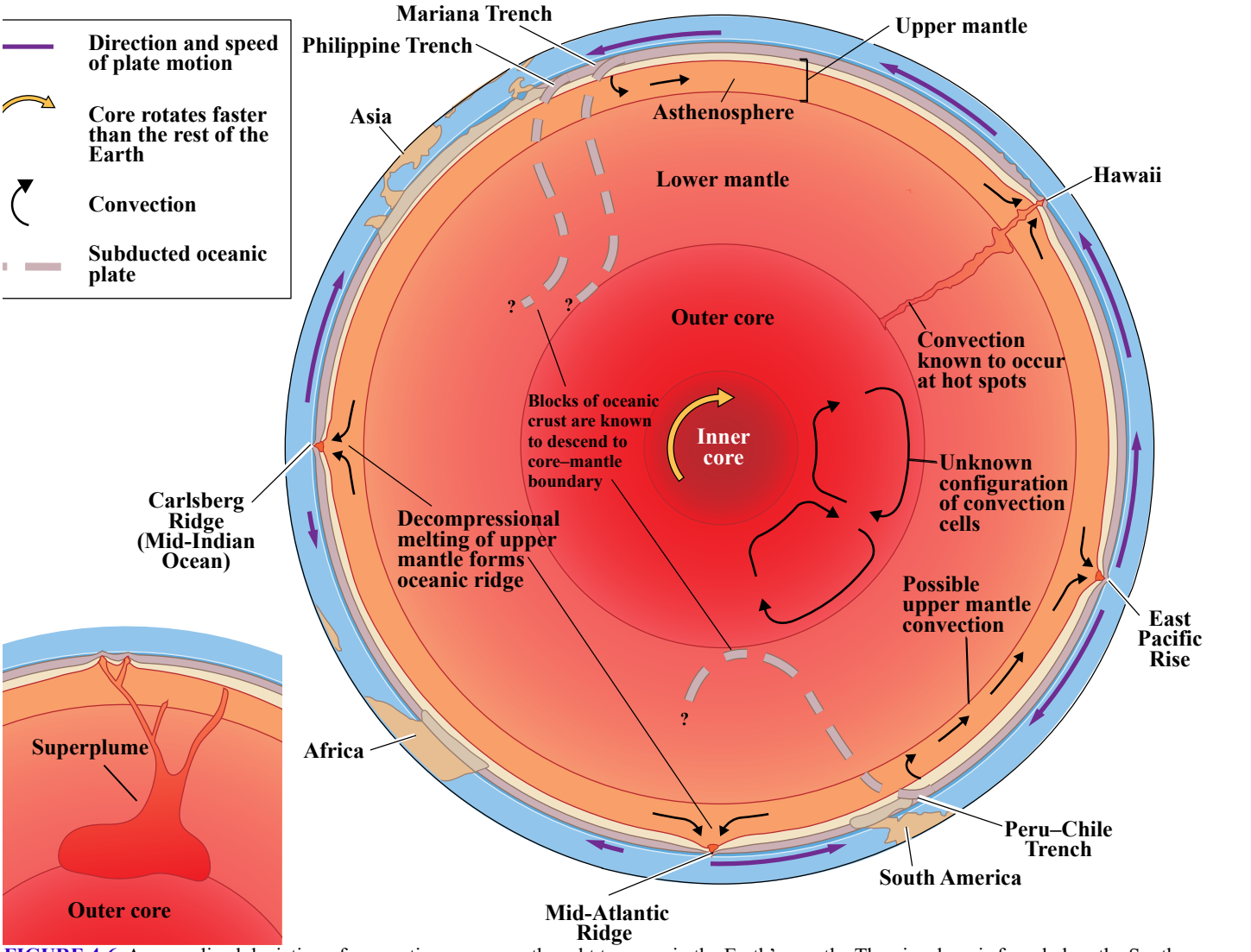

FIGURE 4-6. A generalized depiction of convection processes thought to occur in the Earth's mantle. The view here is from below the Southern Hemisphere. Not quite half of the Earth has been removed by slicing across the Earth at an angle to the equator. This slice removes much of the Southern Hemisphere, including Antarctica, Australia, much of the southern Atlantic and Indian Oceans, and most of the Pacific Ocean as far north as Hawaii and Indonesia. Look at a globe to see this better. It is known that the convection processes beneath some hot spots extend to the core–mantle boundary, and some subducted blocks of lithosphere may sink to this boundary, but the details of mantle convection circulation are not yet known. Convection cells must exist in the liquid outer core, but almost nothing is known about this circulation. The solid inner core rotates a little faster than the rest of the Earth, making an additional revolution about every 400 years. Scientists believe the faster rotation of the inner core may explain the periodic reversal of the Earth's magnetic field and plate tectonic processes that cause the continents to periodically gather together and then spread apart. However, much research is needed before these processes can be understood. Inset: We know little about superplumes but there are two superplumes at present; in the south Pacific Ocean south of Hawaii (in the part of Earth that is cut off in main figure); and under the African continent (further north than the tip of Africa visible in the main figure). These superplumes are thought to be a "blob" with plumes rising from it as depicted.

observations, such as examination of volcanic rocks, studies of earthquake-energy transmission through the Earth, and studies of meteorites. Meteorites are examined because they are believed to represent the types of material that make up the Earth's core and mantle. Until 2023, all samples of rocks thought to be composed of mantle material were likely chemically and mineralogically altered since no sample had ever been obtained from the mantle itself. However, in 2023, the *JOIDES Resolution* recovered a 1.268 m long drill core of nearly continuous mantle rock from a site on the Mid-Atlantic Ridge, where mantle rocks were exposed. Analysis of this core will continue for at least several years, but it has already revealed new information about mantle processes related to hydrothermal vent activity that occurs along the ridge.

A technique called "seismic **tomography**" enables scientists to use earthquake waves to study the Earth's internal structure in more detail than is possible by other means. It has yielded several intriguing findings about the Earth's interior. For example, features resembling mountains and valleys have been found on the core–mantle transition zone. The mountain-like features extend downward into the molten core and are as tall as those found on the Earth's surface. It has been hypothesized that some of these features may be sediment-like accumulations of impurities that float upward out of the liquid nickel/iron core.

PLATE TECTONICS

Over the geological timescales of the Earth's history, oceans and continents have formed and disappeared to be replaced by others through the movements of the lithospheric plates. We now have a basic understanding of how and why the lithospheric plates move and of how plate tectonics has continually shaped and changed the face of the planet.

Driving Forces of Plate Tectonics

The movements of the lithospheric plates are thought to be driven by heat energy transferred through the mantle by **convection** (CC3). The mantle has areas where its constituent material is **upwelled** and **downwelled**. As discussed later in this chapter, the convection processes in the mantle are complex, probably **chaotic** (CC11), and as yet not well understood. However, we do know that the upper mantle loses some of its heat by conduction outward through the Earth's cooler crust. As the upper material cools and contracts, its density increases. When its density exceeds the density of the mantle material below, the cooled material will tend to sink (downwell). This sinking mantle material is balanced by upwelling mantle material that has been heated by the Earth's core and **radioactive** decay within the mantle itself. As a result, this heated mantle material is slightly less dense than the material through which it rises. The precise nature and locations of the **convection cells** in the Earth are not yet known (Fig. 4-6). However, we believe that the convection process extends from the base of the lithosphere all the way to the core-mantle boundary, almost 3000 km below Earth's surface. We do know that flow of mantle material vertically does not take place at a uniform rate as there are discontinuities in the physical properties (that affect density) of mantle material at some depths, notably at about 650 km. Thus, sinking and rising mantle materials slows and speeds up along its path through the mantle and this causes a back up and lateral spreading of subducting slabs and ascending heated mantle material at various depths between the core-mantle boundary and the lithosphere (Fig. 4-6).

The existence of deep convection all the way down to the core-mantle boundary is known primarily from computer tomography. For example, the plumes of higher-temperature mantle material beneath Hawaii and several similar **hot spots** (discussed later in this chapter, Fig. 4-10) have been traced through the entire mantle to the core-mantle transition. In addition, cool, rigid slabs of lithosphere, which apparently have been **subducted** at deep-sea trenches, have been detected deep in the lower mantle close to the core–mantle boundary. It is believed that the slabs of lithosphere sink into the lower mantle, where they are eventually heated and mixed with mantle material, and that some of this mixed material then rises back into the upper mantle at different locations. A single cycle of cooling, sinking, warming, and rising probably takes several hundred million years.

The areas where downwelling is thought to occur in the mantle are beneath the deep-ocean trenches, which surround most of the Pacific Ocean, and are found less extensively in other oceans (Fig. 3-4, Fig. 4-10). The areas under which mantle upwelling is thought to occur include large low shear velocity provinces (LLSVPs). In these areas, **superplumes** the size of continents "swell" more than 100 times Mount Everest's height upward from the core–mantle boundary into the mid-mantle. Two such areas are thought to lie under much of Africa and a large region of the southwestern and central Pacific Ocean, respectively (Fig. 4-6). There are also a number of volcanic hot spots in the crust, some of which are the locations of upwelling from deep within the mantle, and some of which may originate from superplumes. As we will see later in this chapter, there are also other hot spots and areas along the oceanic ridges where magma upwells through the lithosphere. However, this magma is thought to originate from shallow depths within the asthenosphere or upper mantle and be formed by melting of upper mantle material.

Oceanic ridges (also called "mid-ocean ridges") mark the boundaries between two tectonic plates that are moving apart. At these oceanic ridges, the lithosphere is pulled apart as the rigid tectonic plates are moved in response to subduction at other boundaries of each plate. As the plate edges separate, the weight of the lithosphere on the upper mantle at the ridge is reduced, which reduces the pressure on the upper mantle below. That upper mantle material is already hot and close to its melting point. Reduced pressure causes the melting point of the mantle material to become lower, and the upper mantle material melts and rises to fill the gap between the two plates. This results in volcanic eruptions along the ridge crest. The erupted material is cooled and solidifies as it erupts at the seafloor, and this solid material is added to the edges of both plates as new oceanic crust.

The forces that actually move the lithospheric plates across the Earth's surface act in response to the mantle convection, but they are far from fully understood. The plates are believed to be dragged across the mantle's surface as slabs of old, cold, dense lithosphere at the plate edges sink at the deep-sea trenches. The deep-sea trenches mark boundaries along which different plates meet and are called **subduction zones**. The mechanism that moves the plate is somewhat like that of an anchor (the old cold dense subducting edge of the plate) that, when dropped into the water, will drag down the line attached to it (the remainder of the rigid plate), even if, by itself, this line would float. The descending slabs of lithosphere create an additional effect similar to the vortex created as a ship sinks, which drags floating materials down with it. Finally, the plates may also move in response to gravity. Oceanic crust near the oceanic ridges, where new crust is formed, is less dense and floats higher than older oceanic crust. The oceanic crust cools, becomes more dense, and sinks as it moves away from the oceanic ridges toward the subduction zones. Thus, the oceanic crust on each plate lies on a slope, albeit a very shallow one, between the oceanic ridges and subduction zones. The plate thus has a tendency to "slide" downhill (in response to gravity), with newer, higher crust "pushing" older, lower crust down the slope.

Present-Day Lithospheric Plates

Seven major lithospheric plates are generally recognized—the Pacific, Eurasian, African, North American, South American, Indo-Australian, and Antarctic Plates—as well as several smaller plates (Fig. 4-7). However, some plate boundaries are not yet fully defined, and additional small plates that currently are considered to be parts of larger plates undoubtedly exist.

All plates are in motion and, over geological time, both the direction and speed of movement can vary. Figure 4-7 shows the directions in which the plates are moving relative to each other at present. The movement of the plates is extremely slow, just a few centimeters per year, or about the rate at which human fingernails grow. Despite their slowness, plate movements can completely alter the face of the planet in a few tens of millions of years. As they move, plates collide, separate, or slide past each other, and sometimes fracture to form smaller plates. At plate edges, the motion is not smooth and continuous; it occurs largely by a series of short, sharp movements that we feel as earthquakes. Most earthquakes occur at plate boundaries (Fig. 4-8).

Interactions between the moving plates are responsible for most of the world's topographic features. Most major mountain ranges, both on land and undersea, and all the deep-ocean trenches are aligned along the edges of plates. How plate movements

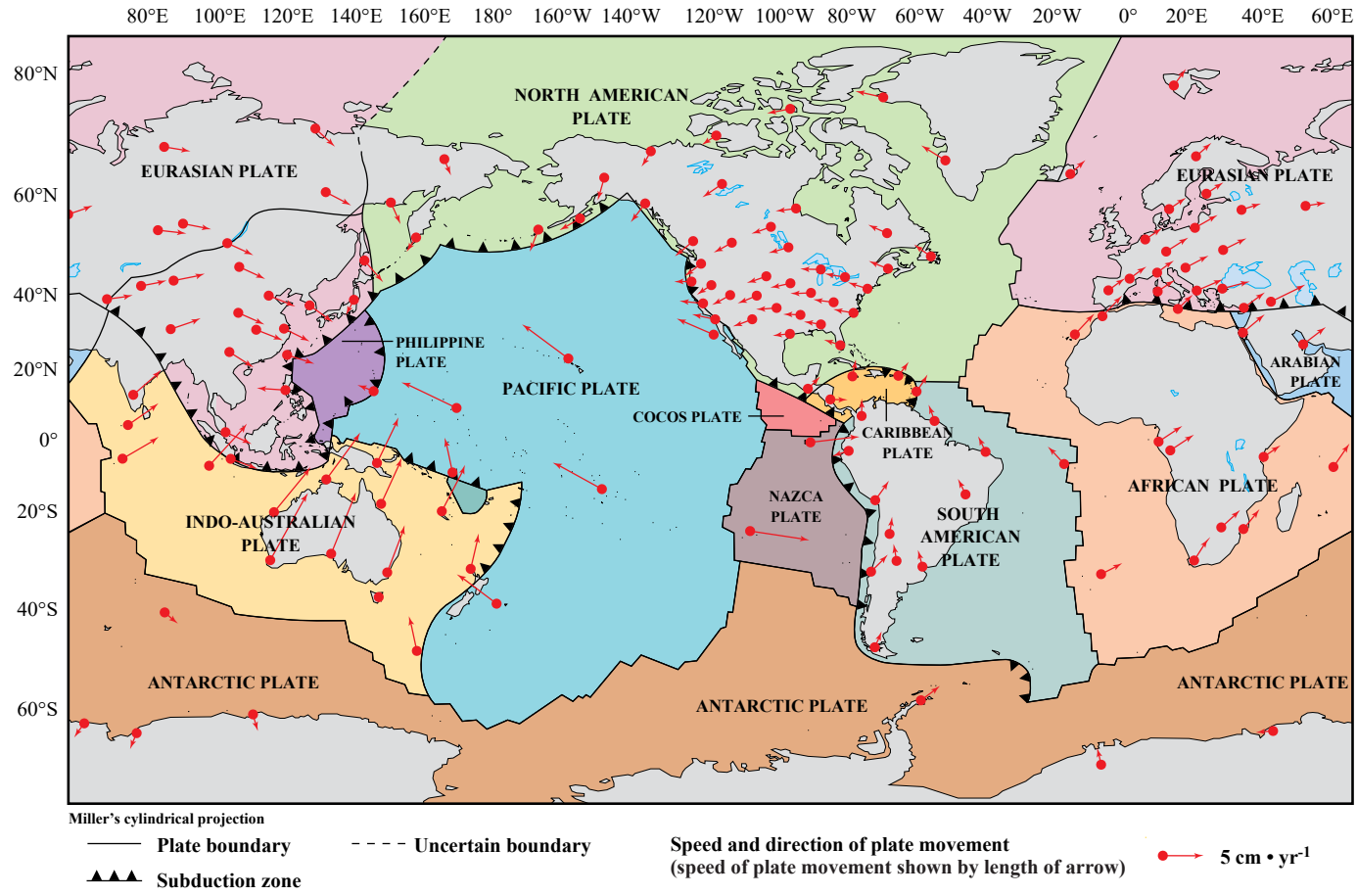

FIGURE 4-7. The major lithospheric plates and their current directions of motion relative to each other. Continental North America lies on the North American Plate, which extends to the center of the Atlantic Ocean. At the Pacific coast boundary of this plate, the North American and Pacific Plates are sliding past each other. The directions of motion show that the Atlantic Ocean is becoming wider, while the Pacific Ocean is becoming smaller.

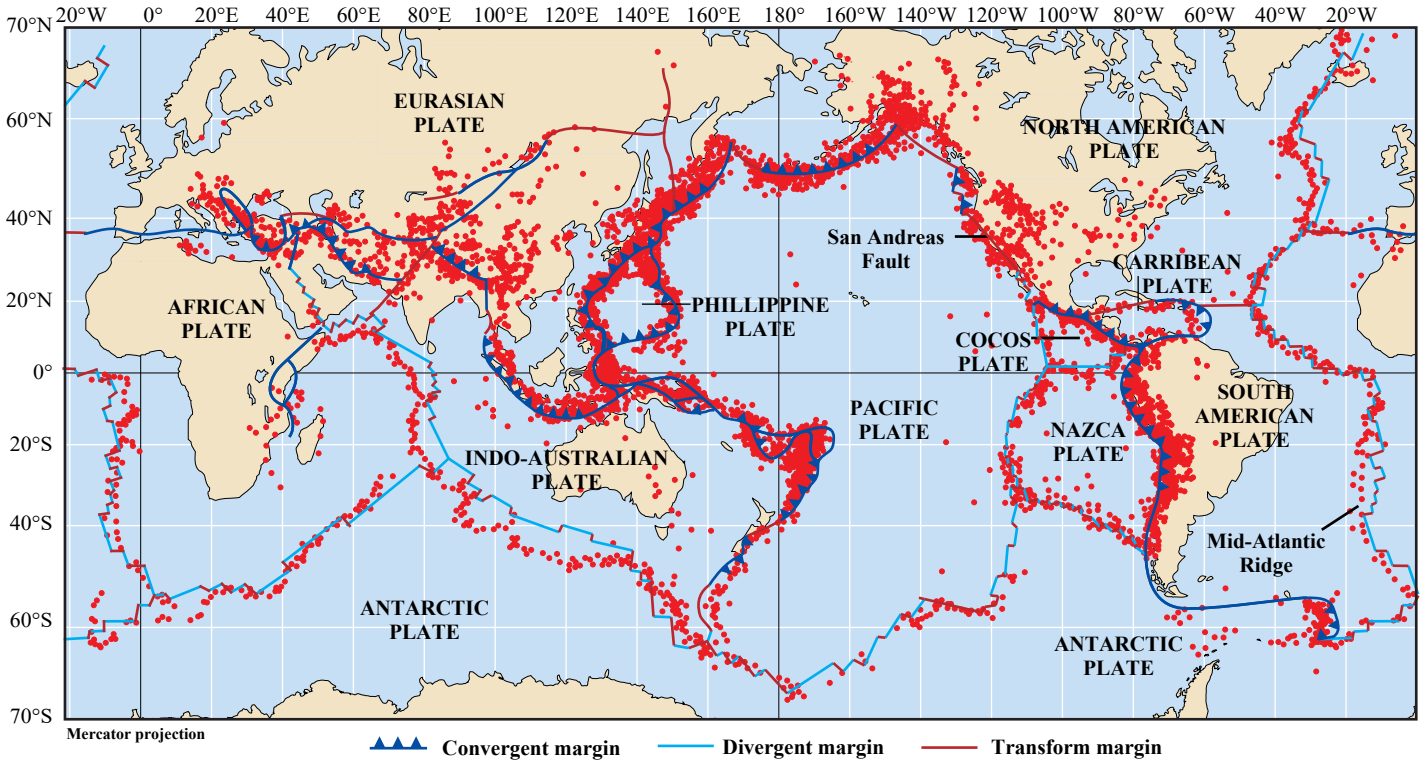

FIGURE 4-8. Locations of the world's earthquakes. Compare this figure with Figure 4-7. Earthquakes are concentrated along plate boundaries and appear to be more frequent at the plate boundaries where plates are converging.

create topographic features is examined later in the chapter.

Spreading Cycles

About 225 million years ago, most continental crust on the Earth was part of one supercontinent called "Pangaea." Since then, Pangaea has broken up and the fragments have spread apart to their present locations. Continental crust breaks apart from a supercontinent, spreads out, and then rejoins to form a supercontinent in a process called a **spreading cycle**. The relative motions of the fragments of Pangaea during the past 225 million years, the current spreading cycle, have been investigated by studies of the magnetism and chemistry of rocks, **fossils**, and ocean sediments, which have revealed subtle clues as to when and where on the planet they were formed. From such studies we can determine from where and how fast the continents have drifted.

The history of the current spreading cycle is quite well known (**Fig. 4-9**). Initially the landmasses now known as Eurasia and North America broke away from Pangaea as a single block. North America later broke away from Eurasia, and South America and Africa separated from Antarctica, Australia, and India. Only much later were North and South America joined, as were Africa and Eurasia. Following the initial breakup of Pangaea, India and later Australia broke away from Antarctica. Since its break from Antarctica, India has moved northward to collide with Asia. This northward movement has been much faster than the rate of movement of other plates.

Although the history of the continents before the breakup of Pangaea is much less understood, Pangaea is known to have been assembled from a number of smaller continents that came together several hundred million years ago. The largest piece of Pangaea, called "Gondwanaland," appears to have remained intact for more than a billion years. The other pieces of continental crust probably broke apart, spread out, and then re-joined to form a supercontinent several times before the formation of Pangaea. Therefore, we refer to the past 225 million years as the "current **spreading cycle**."

As many as 10 spreading and assembling cycles may have occurred during the Earth's history. Each continent is itself a geological jigsaw puzzle, consisting of pieces assembled and broken apart during previous spreading cycles. Today, for example, many areas in the interior of continents show geological evidence that they were subduction zones during earlier spreading cycles (**Fig. 4-10**). Subduction zones form at plate boundaries where oceanic crust is downwelled into the asthenosphere, so these ancient subduction zones could not have been formed where they now lie in the middle of continents. There are several types of plate boundaries, each with its own characteristic topographic features. The various types are identified in the next section, and each type is then further described in subsequent sections of this chapter.

PLATE BOUNDARIES

One of three actions occurs at each plate boundary: two lithospheric plates collide with each other (**convergent plate boundary**), pull away from each other (**divergent plate boundary**), or slide past each other (**transform plate boundary**). Each action deforms the Earth's crust in a different way, creating characteristic topographic features. The behavior of the Earth's crust at a plate boundary depends on the type of crust at the edge of each of the adjacent plates and the directions of movement of the plates in relation to each other.

There are three types of convergent plate boundaries. In the first type, the edge of one **converging** plate is oceanic crust and the edge of the other is continental crust (e.g., the Pacific **coast** of South America); in the second type, both plate edges have oceanic crust (e.g., the Aleutian and Indonesian island arcs); and in the third type, both plate edges have continental crust (e.g., the Himalaya Mountains that divide India from the rest of Asia). At the first two types, where at least one of the plates has oceanic crust on its converging edge, oceanic crust is pushed under, or subducted, into the mantle (**CC3**, **Fig. 4-6**). These are called

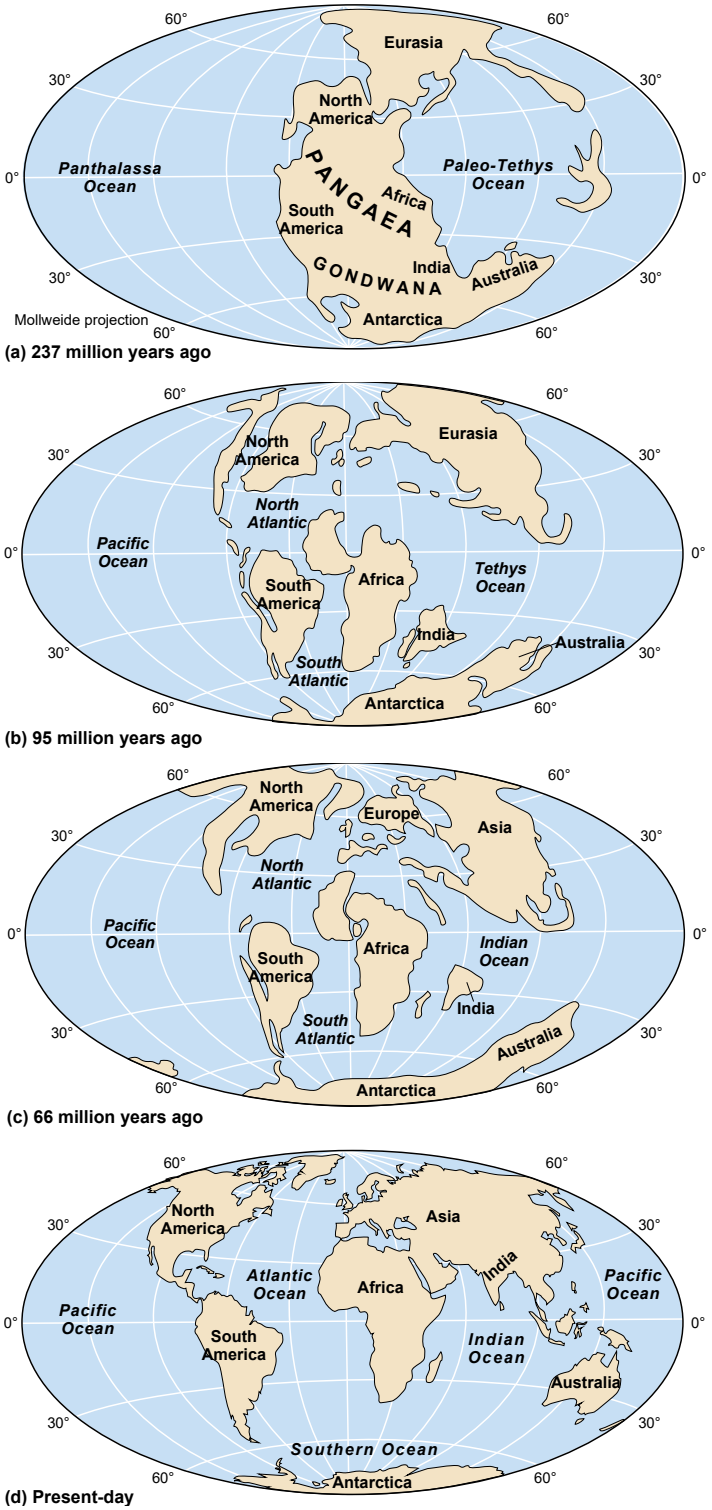

FIGURE 4-9. A reconstruction of the movements of the continents over the past 225 million years, which is the length of the present spreading cycle.

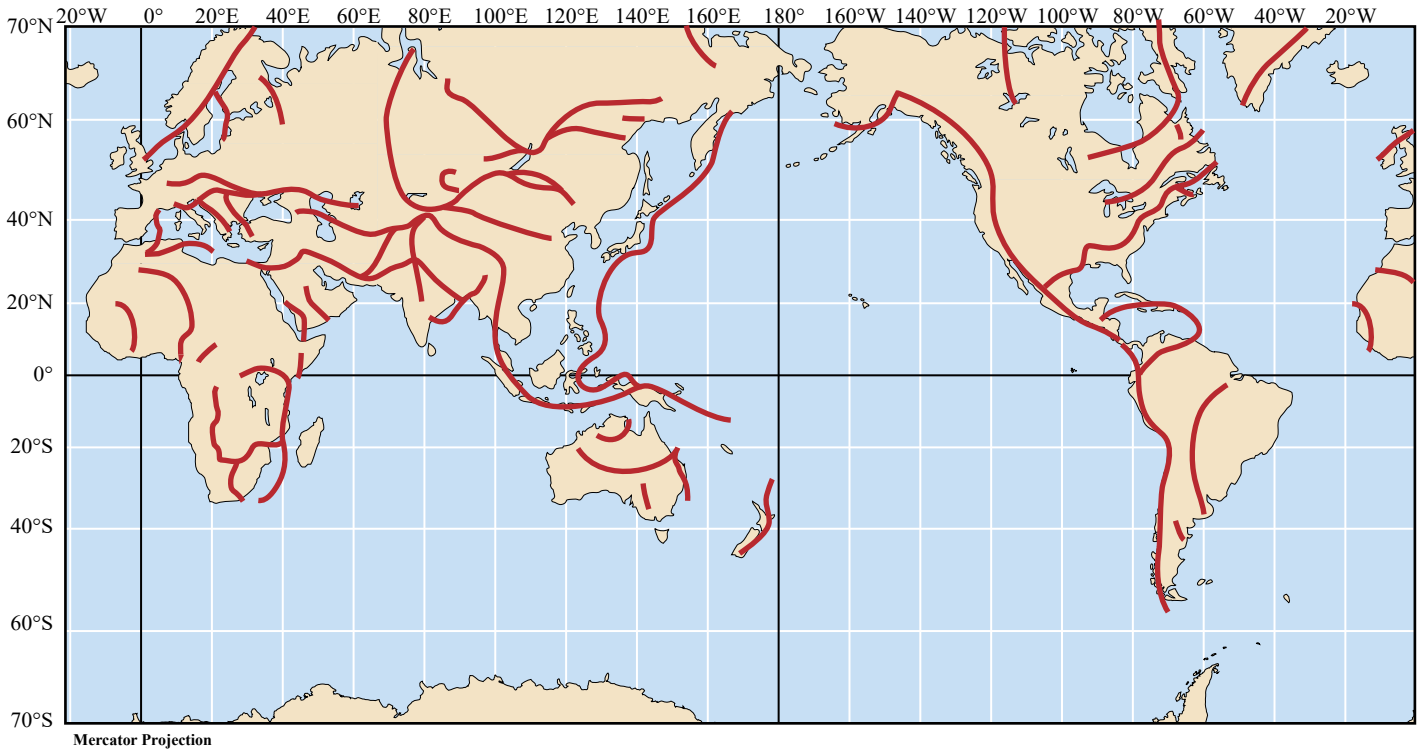

FIGURE 4-10. Sutures and active subduction zones. Sutures are the remnants of subduction zones where plates once converged. The oldest sutures that have been identified are in the interior of continents and are more than 570 million years old.

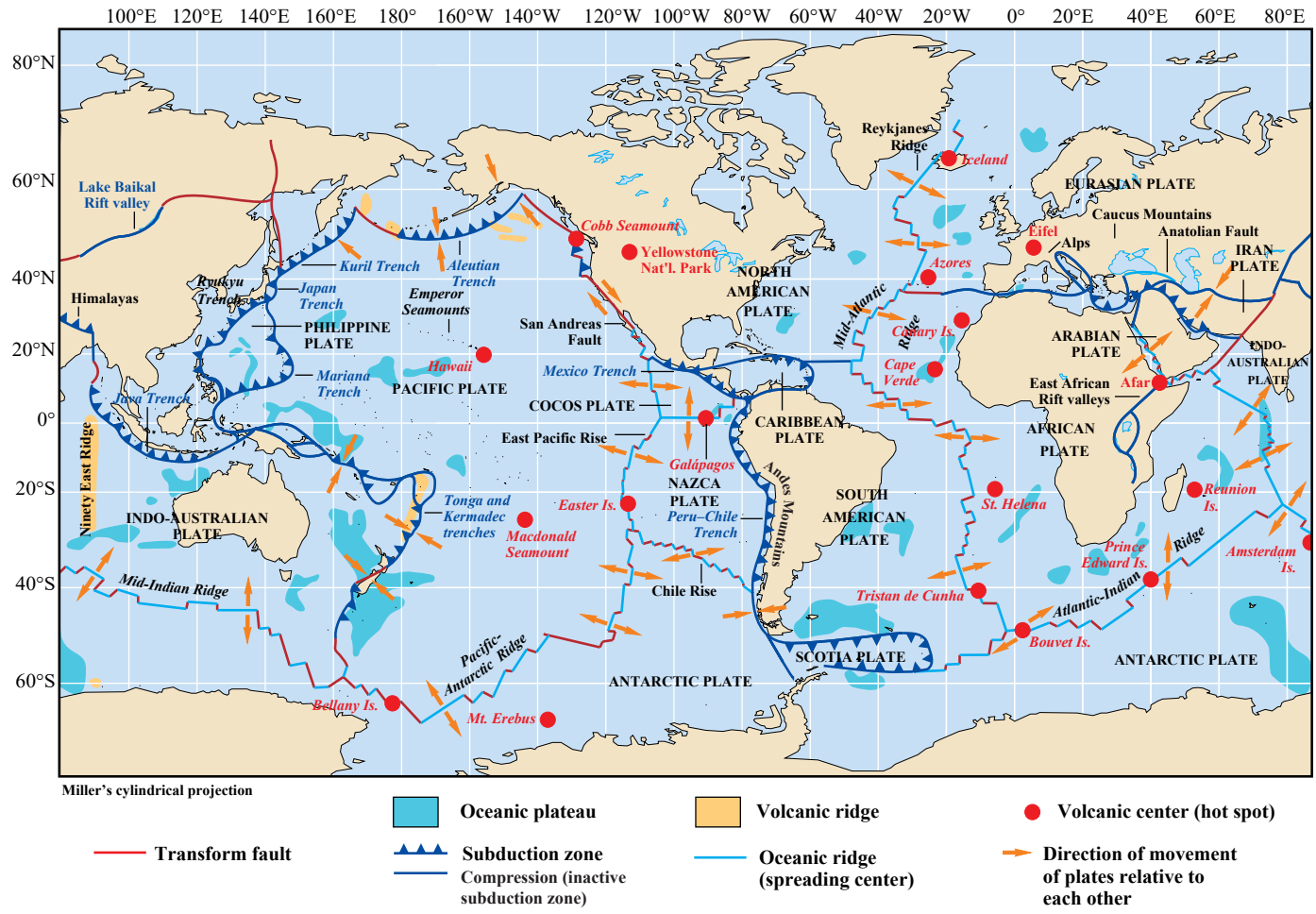

FIGURE 4-11. The locations of hot spots, subduction zones (trenches), and oceanic ridges. Hot spots occur both at plate boundaries and within plates. Subduction zones occur mostly in the Pacific Ocean, and oceanic ridges are interconnected throughout all of the world's oceans.

subduction zones.

Divergent plate boundaries are locations where lithospheric plates are moving apart. There are two types. The first type, called an **oceanic ridge**, is a plate boundary where both plates have oceanic crust at their edges (e.g., the Mid-Atlantic Ridge and the East Pacific Rise). The second type, called a **rift zone**, is a location where a continent is splitting apart (e.g., the East African Rift Zone). If the **divergence** at a rift zone continues long enough, a new ocean will form between the two now separated sections of the continent. The gap that would otherwise be created as two plates move apart at a divergent plate boundary is filled by magma upwelling from below.

Plate boundaries where plates slide past each other are called transform boundaries. There are few long sections of transform plate boundary on the present-day Earth. The best known stretch of transform boundary is on the Pacific coast of North America, primarily California, where the Pacific Plate slides past the North American Plate. Much shorter sections of transform plate boundary occur at intervals along all plate boundaries.

At certain locations along plate boundaries, three plates meet (**Fig. 4-7**) in areas called "triple junctions." There are two types of triple junction: stable and unstable. Stable triple junctions can persist for long periods of time, although their locations may migrate. Unstable triple junctions are those in which the relative motions of the plates cannot be sustained over time because of their geometry. The reason for this is a little complicated, but for a triple junction to be stable, the geometry must be such that each plate can continue to move at the same rate and in the same direction as a whole plate (the plate cannot "bend"). However, the rate at which the plate is subducted or added to by oceanic ridge spreading can be different along its boundaries with each of the other two plates it contacts. Triple junctions where three divergent plate boundaries intersect, as they do near Easter Island in the Pacific Ocean, are always stable. Most other triple junctions are unstable. Unstable triple junctions exhibit enhanced and complex tectonic activity, and their locations move as the interacting plates move relative to each other. An example of an unstable triple junction is the junction where three convergent plate boundaries (the Ryukyu, Japan, and Mariana trenches) intersect south of Japan (**Fig. 4-11**).

CONVERGENT PLATE BOUNDARIES

At two of the three types of convergent plate boundaries, lithosphere is downwelled (subducted) at a subduction zone. Lithosphere (sediments, oceanic crust, and solid upper mantle layer) that enters a subduction zone was formed at an oceanic ridge, generally millions of years earlier (up to about 170 million years ago). Lithosphere slowly cools during the millions of years it travels horizontally at the top of the mantle away from where it was formed. As it cools, its density increases so that, by the time it enters a subduction zone, its density exceeds that of the mantle material beneath it and it can sink into (and through) the asthenosphere.

Why does the lithosphere sink through the asthenosphere only at subduction zones where two plates meet? It would seem that it should sink sooner, since, before it reaches the subduction zone, its density is already higher than that of the underlying asthenosphere. There are two reasons it does not sink sooner. First, the difference in density between lithosphere supporting old, cool oceanic crust and the warmer asthenosphere below is very small. Second, the resistance to flow is high because the lithosphere material is stretched out as a flat plate across the top of the asthenosphere (like a canoe paddle that meets strong water resistance when held flat for the thrust stroke but slices through the water easily when turned sideways). In a subduction zone, the edge of the subducting plate is bent downward (equivalent to turning the paddle blade sideways), thus allowing it to sink much more easily.

On today's Earth, most subduction zones surround the Pacific Ocean (**Fig. 4-11**). At each of these subduction zones, it is the Pacific Plate that is being subducted. The Pacific Ocean floor is being destroyed by subduction faster than new seafloor is created at its ridges, so the Pacific Ocean is becoming smaller. In contrast, the Atlantic and Indian Oceans are becoming larger. The oldest remaining seafloor sediment found in the Pacific Ocean (170 million years old) was retrieved from a hole drilled south of Japan by the International Ocean Drilling Program. All Pacific Ocean oceanic crust that existed prior to 170 million years ago, and the associated sediment, are believed to have entered subduction zones where they were either destroyed or added to the edges of continents. Most of the ocean floor is younger than the Pacific Ocean so almost no oceanic crust older than 170 million years old still exists on Earth because almost all older crust has entered subduction zones and been destroyed. The oldest oceanic crust so far discovered is in a small area called the Herodotus Basin between Cyprus, Crete and Egypt in the Mediterranean Sea dating to about 340 million years old. This area is thought to be a

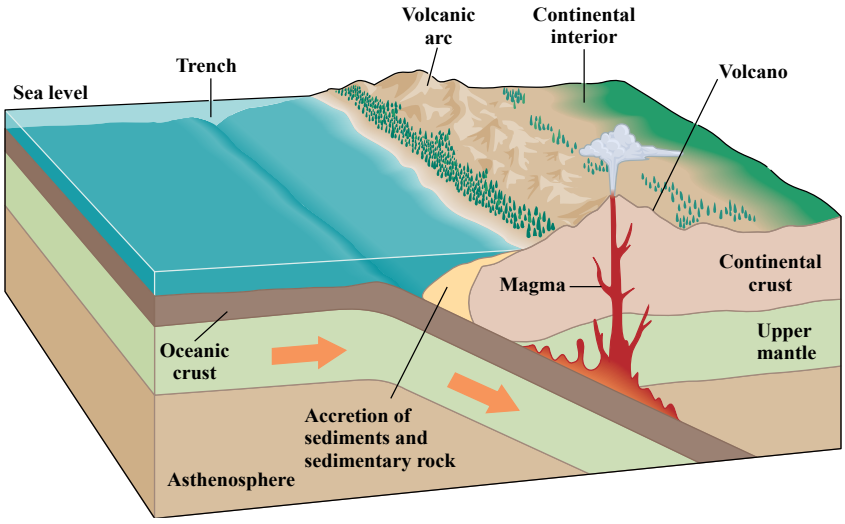

FIGURE 4-12. Subduction zone at a continental margin. A plate with oceanic crust is subducted beneath a plate with continental crust. This process forms a trench, a chain of volcanoes near the edge of the continent, and a chain of coastal hills or mountains, which are formed by compaction of the continental crust edge and accretion of compacted sediments and sedimentary rock from the subducting plate.

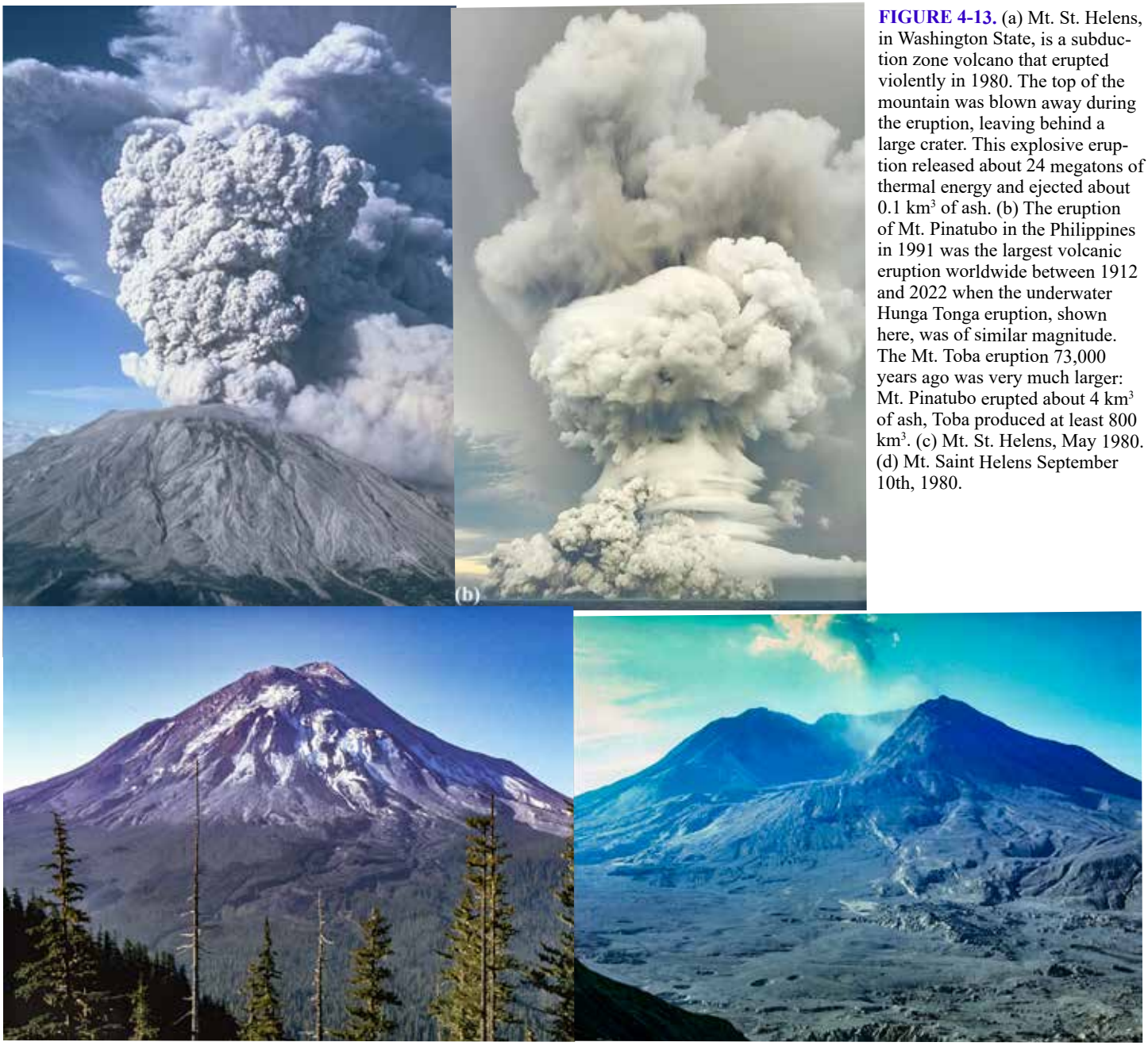

FIGURE 4-13. (a) Mt. St. Helens, in Washington State, is a subduction zone volcano that erupted violently in 1980. The top of the mountain was blown away during the eruption, leaving behind a large crater. This explosive eruption released about 24 megatons of thermal energy and ejected about 0.1 km^3 of ash. (b) The eruption of Mt. Pinatubo in the Philippines in 1991 was the largest volcanic eruption worldwide between 1912 and 2022 when the underwater Hunga Tonga eruption, shown here, was of similar magnitude. The Mt. Toba eruption 73,000 years ago was very much larger: Mt. Pinatubo erupted about 4 km^3 of ash, Toba produced at least 800 km^3. (c) Mt. St. Helens, May 1980. (d) Mt. Saint Helens September 10th, 1980.

fragment of the Tethys Sea that once bordered the supercontinent Pangea (**Fig. 4-9a**).

Subduction Zones Next to Continents

The subduction zones that are located along the coastlines of continents mark convergent plate boundaries where a plate that has oceanic crust at its edge converges with a plate that has continental crust at its edge (**Fig. 4-12**). As the two plates move toward each other, the lithosphere of the plate supporting oceanic crust is thrust beneath the plate that supports the less dense continental crust. As the oceanic crust sinks beneath the continental crust, a subduction zone is formed with a characteristic deep trench parallel to the coast.

The edge of a continent that is carried into the plate boundary at a subduction zone is squeezed and thickened, forming a chain of coastal mountains along the edge of the continent. Because ocean sediments and sedimentary rocks have lower densities, like the continental rocks from which much of the sediment originated (**Chap. 8**), they are relatively buoyant. As a result, these materials tend to be "scraped off" the oceanic plate as they are dragged downward into the subduction zone rather than subducted with it. The sedimentary materials are further compressed, folded, and lifted as more such material accumulates from continuing subduction of additional oceanic crust. Thus, some of the marine sediments from the subducting oceanic crust are collected in the subduction zone and contribute to the formation of the coastal mountains. This process explains why Darwin found marine fossils high in the Andes Mountains (**Chap. 2**).

Volcanoes

Another process that occurs at subduction zones is the formation of a line of volcanoes on the continental crust located a few tens or hundreds of kilometers inland from the plate boundary. The volcanoes are formed when oceanic crust and some associated sedimentary material are subducted beneath the edge of the continental plate. As this oceanic crust and sediment are subducted deeper into the Earth, they are heated by the **friction** of their movement, by the hotter mantle material below, and by increas-

ing pressure. This causes water and other volatile constituents to be released by the subducting plate and its associated sedimentary material as fluids, which migrate upwards and combine with the mantle material above the plate. This causes physical and chemical changes in the mantle material, lowering its melting point, and it melts to form magma with a high concentration of water and other volatile constituents. The resulting magma rises through the overlying continental crust and toward Earth's surface, where it erupts to form volcanoes near the edge of the continental plate (Fig. 4-13a,c,d). The erupting magma is rich in silica, which makes it more viscous (resistant to flow). As the magma erupts, pressure decreases, and water and other volatile constituents become gaseous and expand rapidly. The result is that these eruptions are often explosive, ejecting large amounts of ash and other volcanic material.

Most active subduction zones in which the two plates have continental and oceanic crust at their respective edges are in the Pacific Ocean, and therefore, they are often called "Pacific-type margins." Such margins include the west coast of Central and South America, the west coast of North America from northern California to southern Canada, and the coast of South-central Alaska. These areas have a well-developed ocean trench, coastal mountain range, and inland chain of volcanoes that characterize Pacific-type plate margins (Fig. 4-14).

Exotic Terranes

Throughout the oceans are small areas where the seafloor is raised a kilometer or more above the surrounding oceanic crust. Called **oceanic plateaus**, these areas constitute about 3% of the ocean floor (Fig. 4-11). Some oceanic plateaus are extinct seafloor volcanoes, others are old volcanic ridges, and still others are fragments of continental crust called "microcontinents." Small extinct undersea volcanoes can be broken up and subducted with oceanic crust. However, the larger oceanic plateaus are too thick to be subducted. Therefore, when sections of oceanic plateau enter a subduction zone, their rocks become welded onto the edge of the continent to form **exotic terranes** (Fig. 4-15). Exotic terranes may also be formed when islands, including those originally formed at **magmatic arc** subduction zones (see the next section), are carried into a subduction zone where they impact the edge of a continent. Much of the Pacific coast of North America consists of exotic terranes.

Subduction Zones at Magmatic Arcs

At **oceanic convergent plate boundaries** where both plates have oceanic crust at their edges, a chain of volcanoes erupts magma to form an island chain parallel to the subduction zone. Because the most prominent examples of this type of plate boundary on the present-day Earth are curved (arced), these boundaries are called magmatic arcs (or "island arcs"). Processes that occur at such plate boundaries are similar to processes that occur at subduction zones at a continent's edge. However, because no continental crust is present, the oceanic crust of one plate (which has higher density) is subducted beneath the oceanic crust of the other plate (which has lower density). Because oceanic crust cools with age, and density increases with decreasing temperatures, older oceanic crust has a higher density and is subducted beneath younger crust. For example, at the Aleutian Islands plate boundary, the older Pacific Plate is being subducted below the younger lithosphere of the Bering Sea portion of the North American Plate.

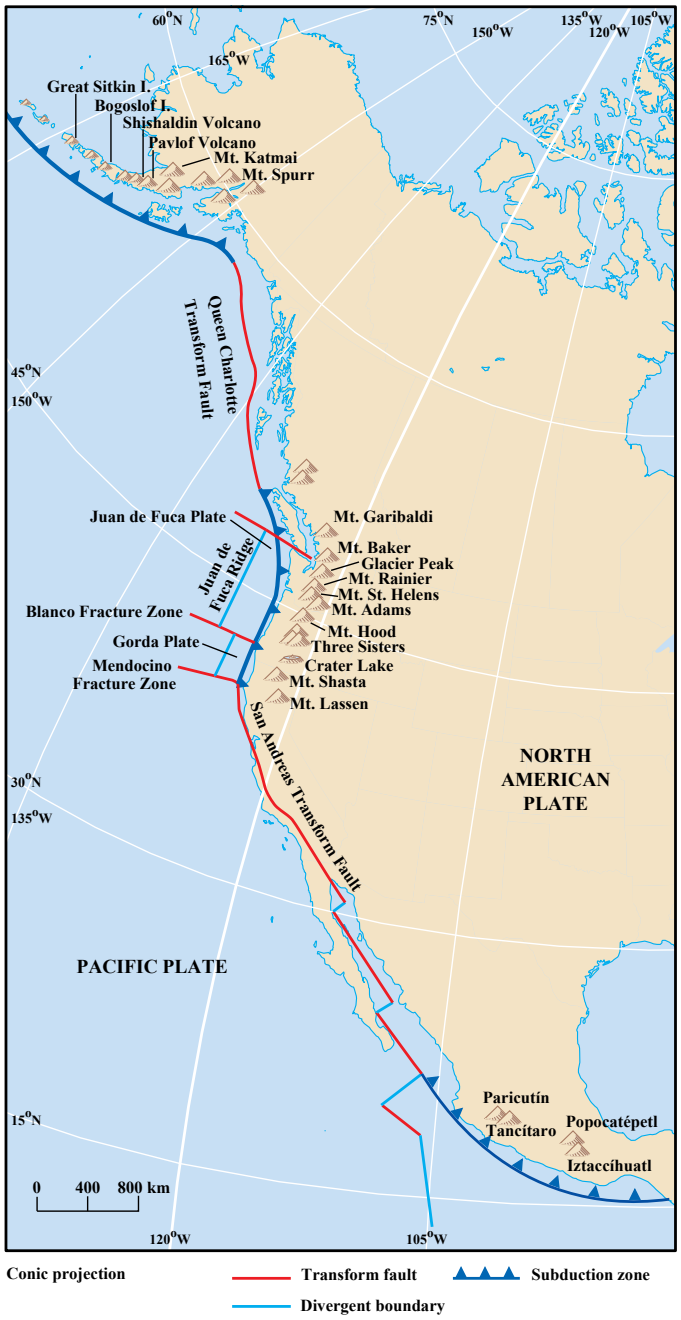

FIGURE 4-14. Stretches of the west coasts of Mexico, California, Oregon, Washington, Canada, and Alaska are characterized by trenches and continental subduction zone volcanoes. The active volcanoes associated with the subduction of the Juan de Fuca and Gorda Plates under the North American Plate do not extend farther south than Mount Lassen in northern California. To the south, the San Andreas Fault, which can be traced from Cape Mendocino through southern California and into northern Mexico, is a transform fault and not a subduction zone. Continuing southward, active volcanoes are again found in central Mexico, where the transform faults of California end and the subduction zone between the Cocos Plate and the North American Plate begins.

The subduction of oceanic crust at magmatic arcs forms a trench system that parallels the plate boundary (Fig. 4-16). Sedimentary materials from the subducting plate accumulate on the edge of the nonsubducting plate, just as they do at subduction zones at a continent's edge. They may form a chain of low **sedimentary arc** islands joined by an underwater ridge called an "outer arc ridge." Behind the ridge is an outer arc basin, an area of the non-subducting plate where the crust is affected little by

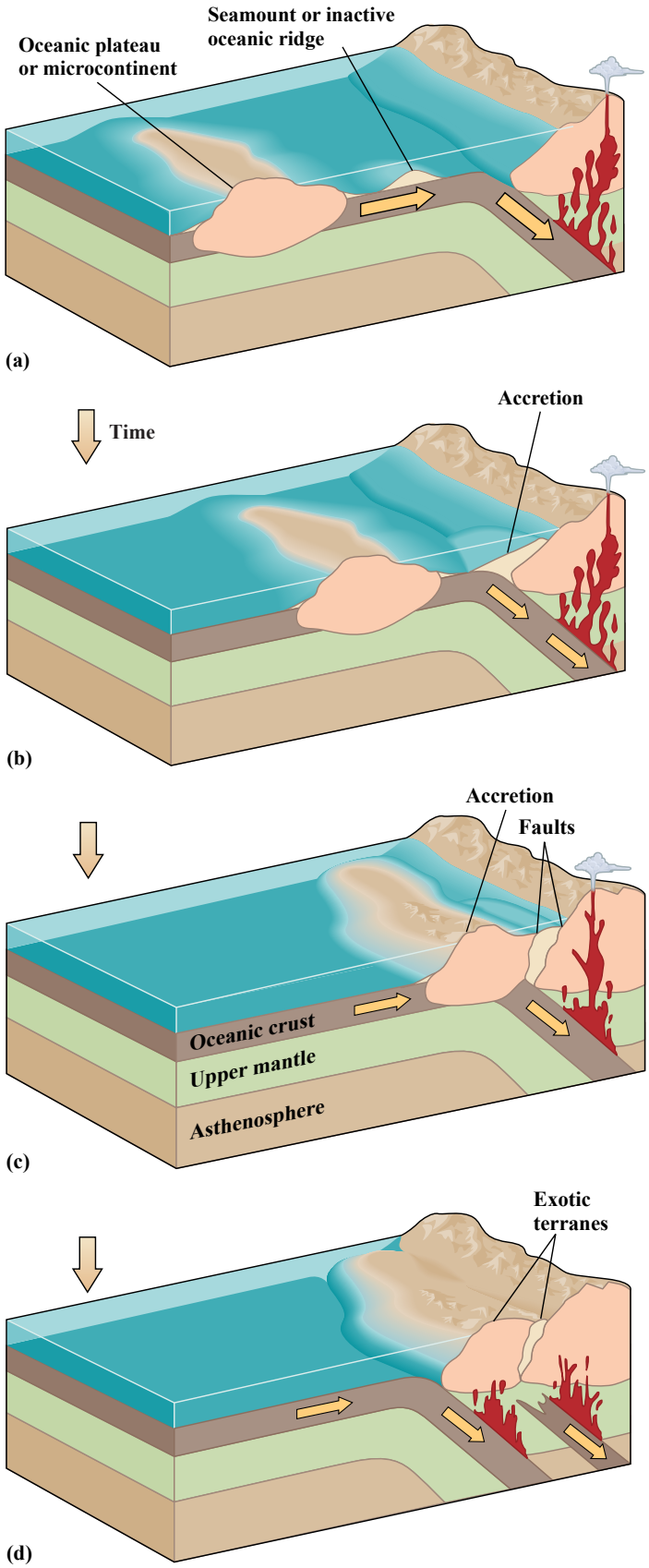

FIGURE 4-15. Formation of exotic terranes. Oceanic plateaus, inactive oceanic ridges, and volcanoes are scraped off the oceanic crust as it enters a subduction zone. The scraped-off material forms new continental crust that is welded onto the edge of the continent and is called an exotic terrain.

the subduction processes.

Most present-day magmatic arc subduction zones are around the Pacific Ocean. They include Indonesia (Fig. 4-17), the Mariana Islands, the Aleutian Islands, and Japan (Fig. 4-6). Each of these areas has the characteristic trench and magmatic arc, some have the low outer arc islands, and some have shallow, sediment-filled **back-island basins**, as described in the next section.

Magmatic Arc Volcanoes

On the nonsubducting plate of a magmatic arc subduction zone, a line of volcanic islands (the magmatic arc) forms parallel to the plate boundary (Fig. 4-16). The islands are constructed by rising magma, which is produced by the sinking, heating, and subsequent melting of subducted oceanic crust and mantle material. These volcanic islands are equivalent to the chain of volcanoes formed on the continental crust of subduction zones at the edges of continents. Like their continental counterparts, magmatic arc volcanoes often erupt explosively because subducted sediments and their associated water are also heated and erupted with the magma. One of the best known of these explosive eruptions is the 1883 eruption of Krakatau (Krakatoa) in Indonesia (Fig. 4-17). This eruption altered the Earth's **climate** for several years afterward, and the eruption and resulting **tsunami** killed an estimated 36,000 people. More recently, in 2022, Hunga Tonga, a large volcano on the Tonga subduction zone whose peak was about 150 m underwater at the time, erupted explosively. The force of the eruption was hundreds of times more powerful than the atomic bomb dropped on Hiroshima in World War 2. The volcano ejected an estimated 10 km^3 of rock ash and sediment and created a massive tsunami that caused damage thousands of kilometers away. Hunga Tonga also caused the largest explosive shock wave that has ever been recorded. This shock wave travelled around the world several times. The eruption also ejected a huge amount of water vapor to the atmosphere, equivalent to about 10% of all the water vapor in Earth's atmosphere at any one time, or enough to fill 58,000 swimming pools. Fortunately, the explosion occurred in a sparsely populated area of the Pacific and caused only 6 known deaths.

At all convergent plate boundaries, the distance between the plate boundary and the line of volcanoes is shorter where the subducting plate's angle of descent is greater. Because the lithosphere cools and its density increases as it ages, older lithosphere tends to sink more steeply (faster) into the asthenosphere than younger lithosphere. Thus, the distance between the trench and the volcano chain is less where the subducting lithosphere is older.

Back-Arc Basins

Sometimes a subducted plate's rate of destruction can exceed the rate at which the two plates are moving toward each other, particularly where old lithosphere is sinking steeply into the asthenosphere. Under these circumstances, the edge of the nonsubducted plate is stretched, which causes a thinning of the lithosphere at its edge. The thinning may create a **back-arc basin** (sometimes called a "back-island basin") behind a magmatic arc (Fig. 4-16). In extreme cases, the lithosphere is stretched and thinned so much that magma rises from below to create new oceanic crust in the basin.

Back-arc basins are generally shallow seas with a floor consisting of large quantities of sediment eroded by wind and water from the newly formed mountains of the magmatic arc and from

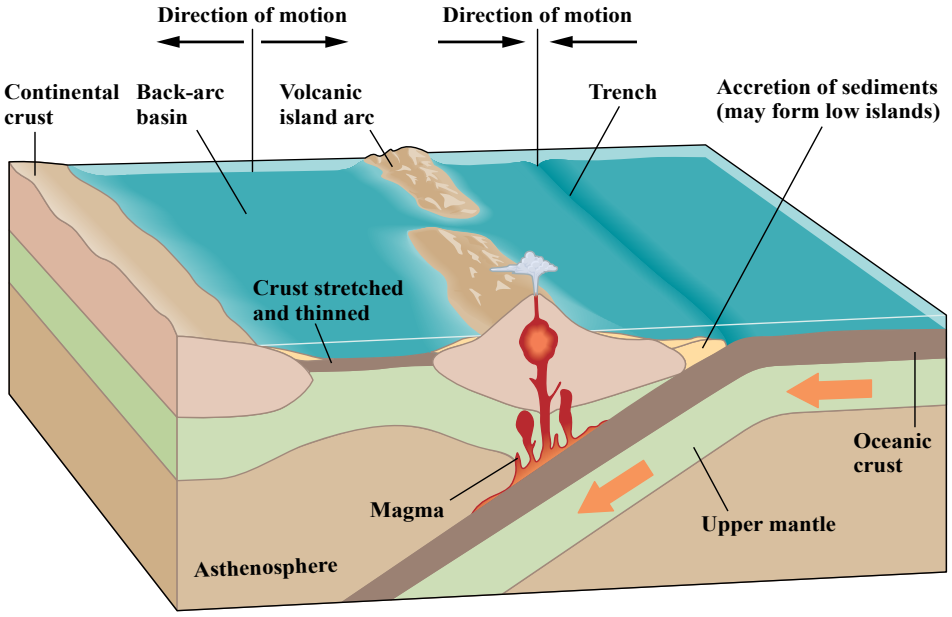

FIGURE 4-16. At magmatic arc subduction zones, as the plate with the higher-density oceanic crust is subducted, a chain of volcanic islands (magmatic or island arc) is formed along the edge of the nonsubducting plate. Sometimes low sedimentary islands also form between the island arc and the trench. At some magmatic arcs, such as shown here, the oceanic plate subducts fast enough that the adjacent plate (shown with a continental crust edge in this example) does not move toward the plate boundary as fast as the subduction occurs. In this situation, a trench and island arc form outside a back-arc basin, which is created by stretching and thinning of the nonsubducting plate edge (the continent edge in this example). The directions of plate motion shown are relative to each other. In some cases, the trench may migrate seaward, while the non-subducting plate migrates more slowly in the same direction so that a back-arc basin is still formed.

nearby continents. The Mariana Trench subduction system, where the Pacific Plate is being subducted beneath the Philippine Plate, provides an example of a back-arc basin. The back-arc basin lies between the Mariana Trench–Island subduction zone and the Philippine Trench. About 200 km to the west of the Mariana subduction zone is a now-inactive oceanic ridge that is the former location of back-arc spreading (Fig. 3-4).

Collisions of Continents

Continental collision plate boundaries form at the convergence of two plates that both have continental crust at their edges. When the lithosphere of one plate meets the edge of the other plate, neither is sufficiently dense to be dragged into the asthenosphere and subducted. Therefore, the continental crusts of the two plates are thrust up against each other. As the collision continues, more continental crust is thrust toward the plate boundary, and the two continents become compressed. One continent is generally thrust beneath the other, lifting it up. The forces created and the energy released by such a collision are truly immense, raising a high mountain chain along the plate boundary. The effect of such collisions is not unlike the effect of a head on collision of two cars, in which each car's hood is compressed and crumpled upward, and one car may ride partially under the other.

Collision plate boundaries are relatively rare on the Earth today, but the geological record indicates that they were more common in the past, such as when Pangaea was formed from preexisting continents. The most prominent continental collision plate boundaries today are the ones between India and Asia and between Africa and Eurasia (Fig. 4-11). Older examples, formed before the present spreading cycle, include the Ural Mountains of Russia and the Appalachian Mountains of North America.

The ongoing India–Asia collision is a particularly vigorous collision. It began about 40 million years ago when the edges of the continental shelves of the two continents first collided (Fig. 4-18). The movement of India toward Asia before the collision was very fast in relation to the speed at which other plates are known to move. The high speed explains the extreme violence of the India–Asia collision, during which India has been thrust under the Asian continent, creating the Himalaya Mountains and compressing and lifting the vast high steppes of Asia. The Himalaya Mountains continue to be uplifted, and the many powerful earthquakes felt throughout the interior of China and Afghanistan are caused by earth movements that release the compressional stress continuously built up by the India–Asia collision. This collision may also be compressing and fracturing the Indo-Australian Plate across its middle, as its northern section is slowed by the collision and its southern section continues northward.

A similar continental collision began about 200 million years ago between Africa and Eurasia. The convergence is slower than the India–Asia convergence, so the collision is less violent. Nevertheless, the Africa–Eurasia collision is responsible for building the Alps and for the numerous earthquakes that occur on the

FIGURE 4-17. The outer arc ridge, volcanoes, and back-arc basins of Indonesia and the Philippines. The numerous active volcanoes on the major islands of Indonesia, including Sumatra and Java, are evidence of the very active nature of this oceanic convergence. The eruptions of Krakatau in 1883 and Tambora in 1815 were two of the largest eruptions of the past several centuries.

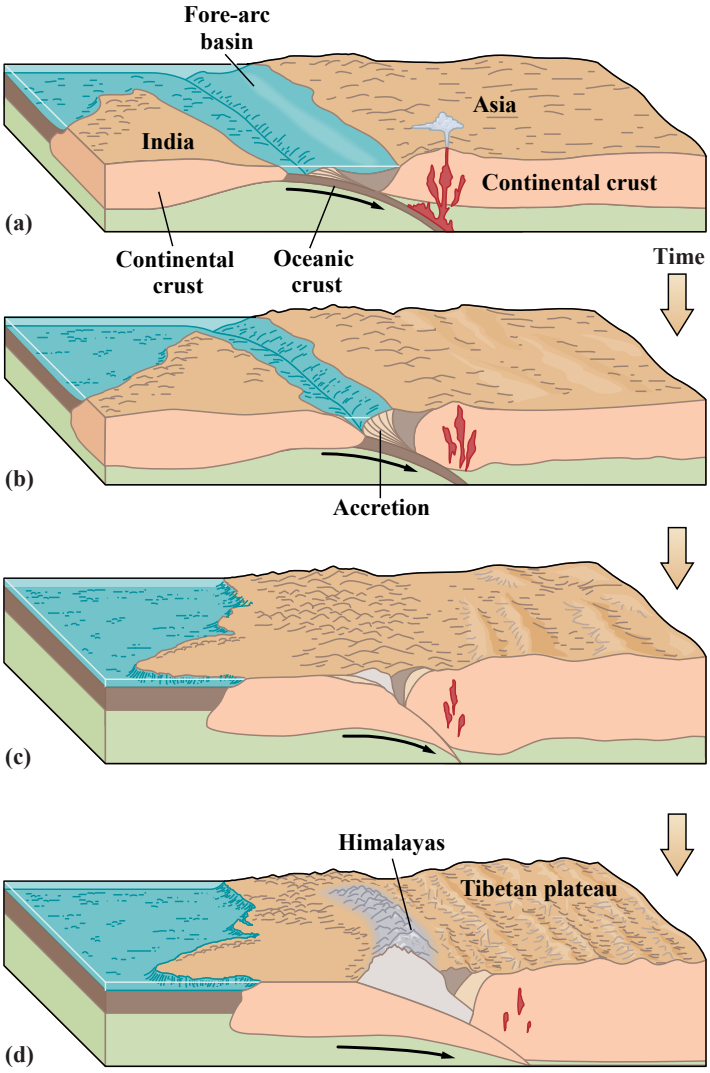

FIGURE 4-18. The continental collision boundary between India and Asia. About 50 million years ago, the northern margin of the Indian continental crust began a collision with the southern margin of the Asian continental crust. India is being thrust under the Asian continent as the collision continues. The Himalayas and the high Tibetan Plateau were both created by the collision. The tops of the Himalaya Mountains are formed from sedimentary rocks scraped off the oceanic crust of the ocean floor destroyed as the continents first came together.

Balkan Peninsula (parts of Turkey, Greece, Albania, Bosnia and Herzegovina, Bulgaria, Croatia, Macedonia, Romania, Slovenia, Serbia, and Montenegro) and the adjacent region of Asia (parts of Turkey, Armenia, Georgia, Azerbaijan, and Iran).

DIVERGENT PLATE BOUNDARIES

The processes by which divergent plate boundaries are first formed are not fully understood. It is believed that rift zones in the continents may be at least partially caused by upwelling of magma accumulated under the continents, and that this magma may accumulate because the continental crust acts as an insulator and causes heat to build up in the mantle below it. However, the most prominent rift zone on the Earth today, the East African Rift Zone, is thought to have been caused by upwelling associated with the superplume that is believed to exist in the lower mantle beneath the African continent.

At oceanic ridges, new oceanic crust is continuously formed by upwelling of magma from the asthenosphere as the plates move apart, but this upwelling is not the cause of the plate divergence. Instead, the magma upwelling occurs in order to fill the gap created when sinking (downwelling) of lithosphere at a subduction zone elsewhere on one (or both) of the plates moves the entire plate toward this subduction zone, "pulling" the plate away from the oceanic ridge and creating a divergent boundary.

Oceanic Ridges

When two plates are pulled apart (diverge), the gap between them is filled by upwelled magma. The reduction of weight on the upper mantle that occurs as the plate edges are drawn apart causes solid mantle material to melt as pressure is reduced. This hot, low-density magma rises through the splitting crust and spreads on the seafloor from a series of volcanoes aligned along the gap. This is the primary process responsible for building the connected undersea mountain chains of the oceans (**Fig. 3-4**) and is often called **seafloor spreading**. The mountain chains are known as oceanic ridges and oceanic rises. Most are in the approximate center of their ocean basins and follow the general shape of the coastlines on either side of the ocean (**Fig. 4-11**). An example is the Mid-Atlantic Ridge, which bisects the Atlantic Ocean from near the North Pole to the southern tip of South America (**Fig. 4-19, Fig. 3-4**).

As the plate edges separate at an oceanic ridge, new oceanic crust is formed by a sequence of processes that cause oceanic crust to be layered. Molten magma upwells from below and forms a chamber of molten magma below the seafloor. Some magma cools and solidifies on the sides of this chamber to form a rock called "gabbro." This becomes the lowest of four layers that are found consistently in almost all oceanic crust. The next layer up is formed by magma that rises through vertical cracks and solidifies in wall-like sheets called "dikes." The third, and initially upper, layer is composed of pillow basalt, which is formed when erupting **lava** is cooled rapidly by seawater. The final layer of oceanic crust is composed of sediments deposited from the ocean water column.

New crust formed at an oceanic ridge or rise is hot and continues to be heated from below as long as it remains above the divergence zone. Because it is hot, it is less dense than older, cooler crust. Beneath the new crust, the solid mantle portion of the lithosphere is also thin. Therefore, the new crust floats with its base higher on the asthenosphere (**CC2**), and consequently, the seafloor is shallower. As the new crust is moved progressively farther from the divergent plate boundary, it cools by conduction of heat to the overlying water and becomes denser. In addition, the lithosphere slowly thickens as mantle material in the asthenosphere just below the base of the lithosphere cools, solidifies, and is added to the lithospheric plate. Therefore, the aging crust on the plate sinks steadily to float lower on the asthenosphere. Thus, oceanic ridge mountains slowly move away from the divergence zone and sink, while new mountains form at the divergence zone to take their place.

Oceanic Ridge Types

There are three basic types of oceanic ridges distinguished by spreading rate, the rate at which the two plates are being pulled apart. Two of these types constitute a majority of the length of the world's oceanic ridge system. The third, where the spreading rate is very slow, is less well studied.

The two faster-spreading types of oceanic ridges have certain common characteristics. Each has a string of volcanoes aligned along the axis of the ridge. The volcanoes create a continuously

FIGURE 4-19. Topography of the Arctic Ocean floor. The Mid-Atlantic Ridge extends. into the Arctic Ocean. This extension is named the Gakkel Ridge. Shallow sills restrict water circulation between the Arctic Ocean and other oceans. The sill between the Arctic Ocean and the Bering Sea is especially shallow, and the circulation between the Arctic and Pacific Oceans is more restricted than the circulation between the Arctic and Atlantic Oceans.

spreading layer of new oceanic crust on the ocean floor. The most common type of oceanic ridge, exemplified by most of the Mid-Atlantic Ridge, has a well-defined, steep-sided **central rift valley** extending down its center (**Fig. 4-20, Fig. 3-4**). The valley is the gap formed by the two plates pulling apart. Numerous earthquakes occur beneath the valley, and volcanic vents erupt magma to create the new oceanic crust. Central rift valleys are generally 25 to 50 km wide and 1 to 2 km lower than the surrounding peaks. The oceanic ridge mountains stand 1 to 3 km above the surrounding ocean floor and are extremely rugged.

The second type of oceanic ridge, exemplified by the East Pacific Rise, is much broader and less rugged than the Mid-Atlantic type, and the central rift valley is absent or poorly defined (**Fig. 3-4**). The lack of a central rift valley is thought to be due to a faster spreading rate. The less rugged topography of this type of ridge is reflected in the seafloor of the surrounding plate, which is flatter than the seafloor surrounding more slowly spreading ridges.

The third type of oceanic ridge occurs where the spreading rate is extremely slow. This type of oceanic ridge has very few and widely spaced volcanoes. In addition, much of the new seafloor generated at these ridges is missing the characteristic basalt layer found elsewhere. This layer is thought to be missing because magma at these slowly spreading ridges has time to cool and solidify before it rises to the seafloor. The entire process appears to be different at these very slowly spreading ridges than at other oceanic ridges. At very slowly spreading ridges, the seafloor is apparently just cracked apart between volcanoes, and warm but solid rock then rises to become new seafloor. The Gakkel Ridge beneath the Arctic Ice Cap, northeast of Greenland, and the Atlantic-Indian Ridge as it snakes around the southern tip of Africa are examples. By some estimates, up to 40% of the oceanic ridge system worldwide may be of this extremely slowly spreading type.

Oceanic Ridge Volcanoes

Where the peaks of oceanic ridge volcanoes approach the ocean surface, as they do near Iceland, eruptions can be violent. In such cases, numerous small steam explosions are created at the volcano vent when hot erupting magma and seawater come in contact (you may have seen video footage of such eruptions). However, most eruptions at oceanic ridge volcanoes are quiet and smooth-flowing, despite the magma–seawater contact. There are several reasons for the lack of explosiveness. First, unlike the magma of subduction zone volcanoes, the magma rising into oceanic ridge volcanoes does not include sediments and water. Second, water movements carry heat away from the erupting magma, minimizing or preventing the production of steam. Third, the boiling point of water is several hundred degrees Celsius at the high pressures present at the depths of most oceanic ridge volcanoes.

Because they are not explosive, and they occur deep within the ocean, most oceanic ridge volcanic eruptions occur without being noticed. Nonetheless, they can be at least as large and as frequent as eruptions on land. For example, a sidescan **sonar** survey of part of the East Pacific Rise in the early 1990s revealed a huge lava field formed by an eruption believed to have occurred within the previous 20 years. The lava field covers an area of 220 km^2 and has an estimated average thickness of 70 m. The volume of lava is estimated to be about 15 km^3— enough to cover 1000 km of four-lane highway to a depth of 600 m, or to repave the entire U.S. interstate highway system 10 times. The lava field is the largest flow known to be generated during human history (much larger lava fields from prehistoric eruptions are known). Only a small fraction of the 60,000 km of oceanic ridge has been studied by sidescan sonar surveys or by **submersibles**. Hence, the huge young lava field on the East Pacific Rise may be dwarfed by undiscovered flows from other undersea eruptions.

In 2009, the ROV *Jason II* captured video of an eruption of the West Mata volcano, located 1100 m deep on the Tonga Ridge. Several more underwater volcano eruptions have now been captured on video and studied extensively by scientists (watch an

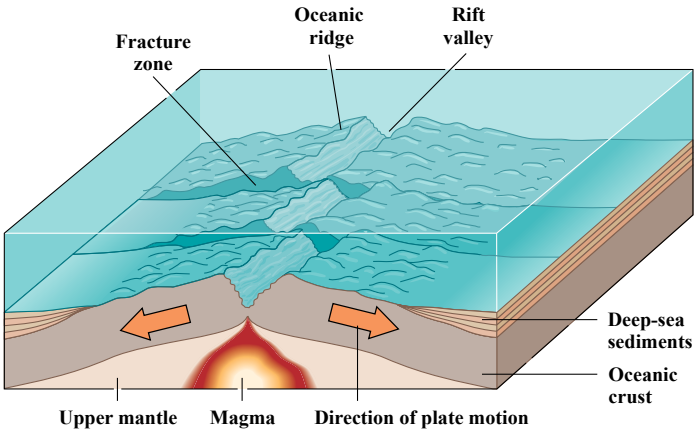

FIGURE 4-20. Oceanic ridge divergent plate boundaries are characterized by a ridge with a central rift valley and by fracture zones. On slowly spreading ridges, such as the Mid-Atlantic Ridge, the rift valley may be 30 km wide and up to 3 km deep, and the ridge is rugged with steep slopes. On rapidly spreading ridges, such as the East Pacific Rise, the rift valley is typically 2 to 10 km wide and only 100 km or so deep, and the ridge is broad with low-angle slopes and relatively smooth topography.

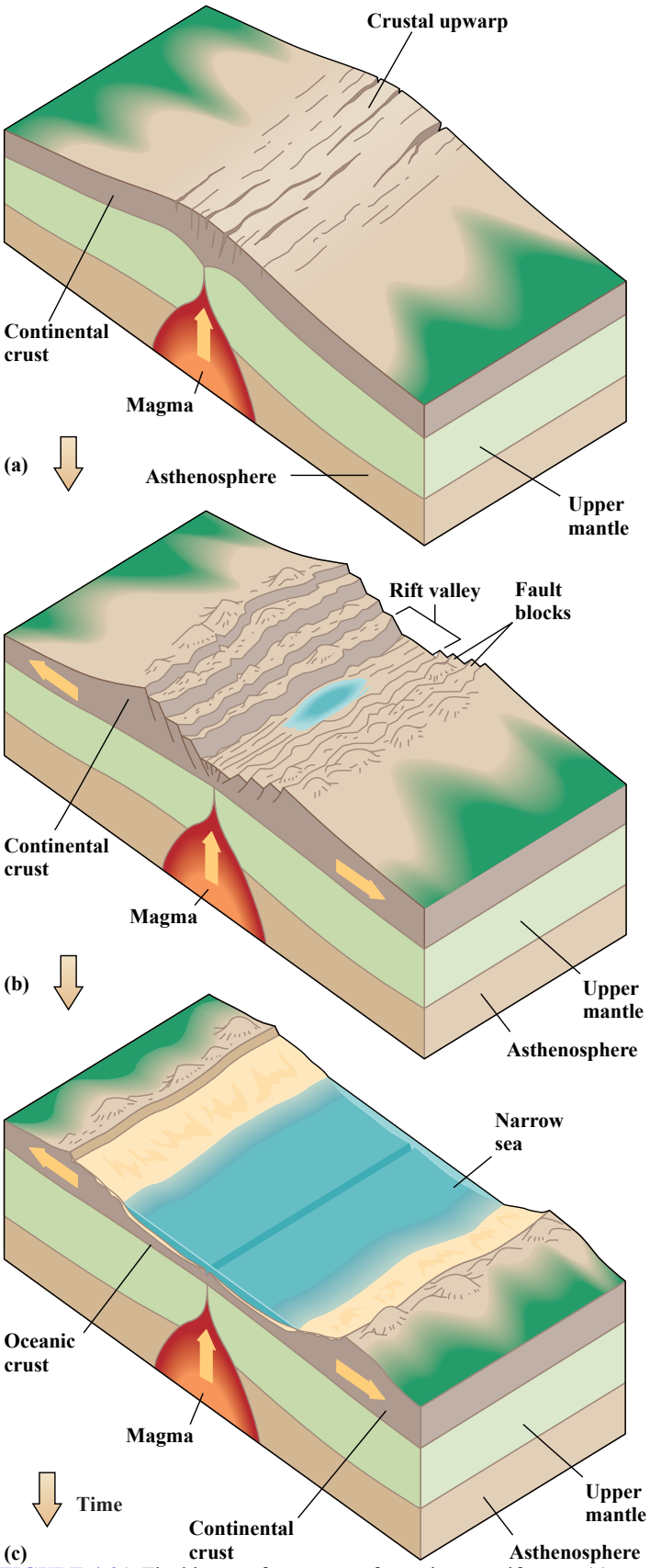

FIGURE 4-21. The history of new ocean formation at a rift zone. (a) Rising magma beneath a continent pushes the crust upward and creates cracks and fractures. (b) The crust is stretched and thinned. Collapses occur along a series of fault blocks, leading to the formation of a rift valley. Lakes may be formed in the rift valley. (c) The rift widens, allowing ocean water to enter and creating a narrow sea, such as the Red Sea. The rift continues to widen, and an oceanic ridge is formed down its center, eventually creating a new ocean.

eruption here). It appears these underwater volcanic eruptions are far more frequent and numerous than volcanic eruptions on land.

Rift Zones

Rift zones occur where a continent is being pulled or is breaking apart. If the rifting continues long enough, an ocean basin is created between two separate, smaller continents. A rift zone may be created when a continental crustal block remains in one place for a prolonged period. In such instances, heat beneath the continent is partially trapped by the insulating effect of the continental rocks. The mantle temperature may increase until the rock at the base of the lithosphere melts, at which time hot magma could rise and split the continental block. Rift zones may also be formed by variations in the convection processes operating within the mantle that cause the mantle upwelling to become located under the continent (**CC3**). As yet, we have no detailed understanding of the relative roles of mantle convection and the insulating effect of continental crust, or of how these processes interact to split continents and create new oceans. In contrast, the processes that occur once a rift zone has been initiated are much better understood.

New Oceans

A new ocean may form in a location where the temperature of the mantle below the continent becomes elevated (**Fig. 4-21**). The process may occur as follows: As the mantle temperature increases, its density decreases and it rises isostatically (**CC2**), causing the crust to be thrust upward to form a dome. The upward thrust stretches the crust, causing fractures that extend outward from the center of the dome. As such a fracture widens, the lithosphere beneath the bulge begins to melt. Eventually, blocks of crust break off and slip down into the rift valley formed by the fractures. This occurs unevenly along the length of the rift, with some sections of the rift valley being deeper than others.

Initially, the rift valley floor may be well above sea level. In wet climates, the deeper areas of the valley may fill with rainwater, forming long, narrow lakes. Volcanoes form in the rift valley and on its sides as magma upwells through cracks left by the fracturing and slipping blocks of crust. The sides of the rift move steadily apart, and magmatic rocks accumulate at the bottom of the valley. Because magmatic rocks, when they cool, are denser than the continental rocks being displaced to the sides of the rift, the rift valley floor sinks until it is eventually below sea level. At this point, the rift may fill with seawater and become an arm of the ocean. However, new volcanoes and landslide debris falling from the rift valley flanks may fill and temporarily increase the elevation of the rift valley floor. Thus, the connection with the ocean may be made and broken numerous times before the rift valley becomes a "permanent" **marginal sea**.

As the rift continues to widen, the original continent splits into two separate continents, separated by a widening ocean with an oceanic ridge at its center. The rift valley provides a route through which the excess heat built up below can be released. If the quantity of heat is limited, the rifting process may stop before an ocean is formed. However, if excess heat is supplied at a high rate, the spreading may continue, and a new ocean may be formed.

Of the several rift zones on the Earth today, the East African Rift is among the newest (**Fig. 3-4**). The African continent appears to have remained at its present location for more than 100 million years. It is elevated several hundred meters higher than

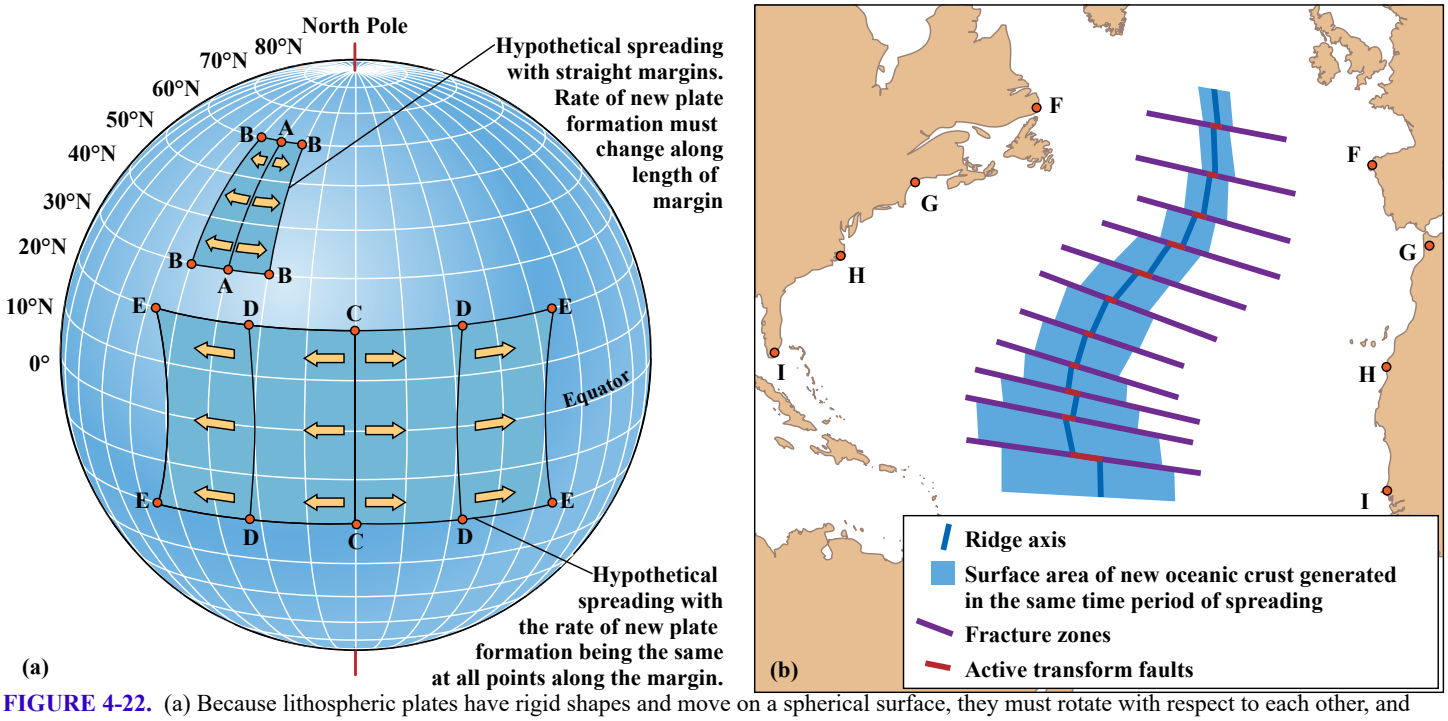

FIGURE 4-22. (a) Because lithospheric plates have rigid shapes and move on a spherical surface, they must rotate with respect to each other, and the rate of formation or destruction of crust at any plate boundary must vary along the plate boundary's length as shown by the spreading from A to B in this figure. The principle is similar to the movement of your eyelids as you open and close your eyes. The eye is exposed or covered more rapidly at the center than at the two corners. The rate of spreading cannot be equal along the plate boundary because the plates would need to bend and deform. This is shown by the vertical lines through the sequence C–D–E , which represent where the plate divergent boundary would be if spreading were uniform along the plate boundary. (b) Each of the points labeled with the same letter on either side of the Atlantic Ocean were joined at one time. Since the pairs are now separated by different distances, crust must have been formed at different rates along the oceanic ridge. Because the lithosphere is rigid and cannot stretch along its edge, variation in the rate of formation or destruction of crust is accommodated by the formation of transform faults between segments of the plate boundary.

most of the other continents and is believed to be directly above a mantle superplume. The East African Rift displays a rift valley, rift valley lakes, and rift volcanoes typical of newly formed continental divergent plate boundaries. The Red Sea is probably a more advanced rift zone, where a connection with the ocean, perhaps a permanent one, has already been made.

TRANSFORM FAULTS AND FRACTURE ZONES

Transform faults are formed where two plates slide past each other. Numerous earthquakes occur along such faults as the edges of the two plates periodically lock and then break loose and slide past each other. The best-known transform fault is the San Andreas Fault in California, where a long section of the Pacific Plate is sliding northward past the North American Plate (**Fig. 4-14**). Most transform faults are much shorter than the San Andreas Fault.

Motions of plates at transform faults do not create the dramatic trenches, volcanoes, and mountain ranges of other types of plate boundaries. However, the movement of the two adjacent plates creates complicated stresses where the two plates meet, produces faults along which earthquakes occur, and forms low hills or mountains. Mountain building may be enhanced at bends on the plate boundary. One such bend on the San Andreas Fault is at the Santa Cruz Mountains south of San Francisco and was the site of the October 1989 Loma Prieta earthquake.

Numerous short transform faults are formed at divergent and convergent plate boundaries. As two plates move apart or together, they also rotate in relation to each other because the plates are moving on a spherical Earth (**Fig. 4-22**). The rate of spreading and creation of new oceanic crust will vary along the plate boundary when such rotation occurs. To accommodate the varying rates of spreading or subduction between two rigid plates, new plate material must be added or destroyed at different rates in adjacent segments of the plate boundary. This is accomplished by the creation of transform faults between short sections of the plate boundary (**Fig. 4-23**). In the transform fault region, the edges of the two plates slide past each other, one side moving slightly faster in one direction than the other, which is moving in the opposite direction.

Transform faults not only accommodate the changing rates of plate creation or destruction but also affect seafloor topography. To examine how transform faults work and how they influence topography consider the oceanic ridge transform fault depicted in **Figure 4-23a**. The transform fault is the segment of the plate boundary where the line of the oceanic ridge is offset. Along this segment, the sections of crust newly formed on each of the two plates slide past each other. However, once the spreading motion has transported the new oceanic crust past the central rift valley of the adjacent segment, it is no longer adjacent to the other plate. Instead, it is adjacent to another piece of its own plate. Because the two sides of this joint are now parts of the same plate and moving in the same direction, the two sides do not slide past each other. These areas are called **fracture zones**. The topographic roughness formed at the transform fault remains after the fracture zone has moved away from the plate boundary. **Figure 3-4** shows numerous transform faults and associated fracture zones that cut perpendicularly through the oceanic ridges. At each transform fault, the line of the oceanic ridge crest is offset.

Some transform faults connect divergent and convergent plate boundaries, and others occur in subduction zones. The seafloor at subduction zones is covered by deep sediments that flow into any

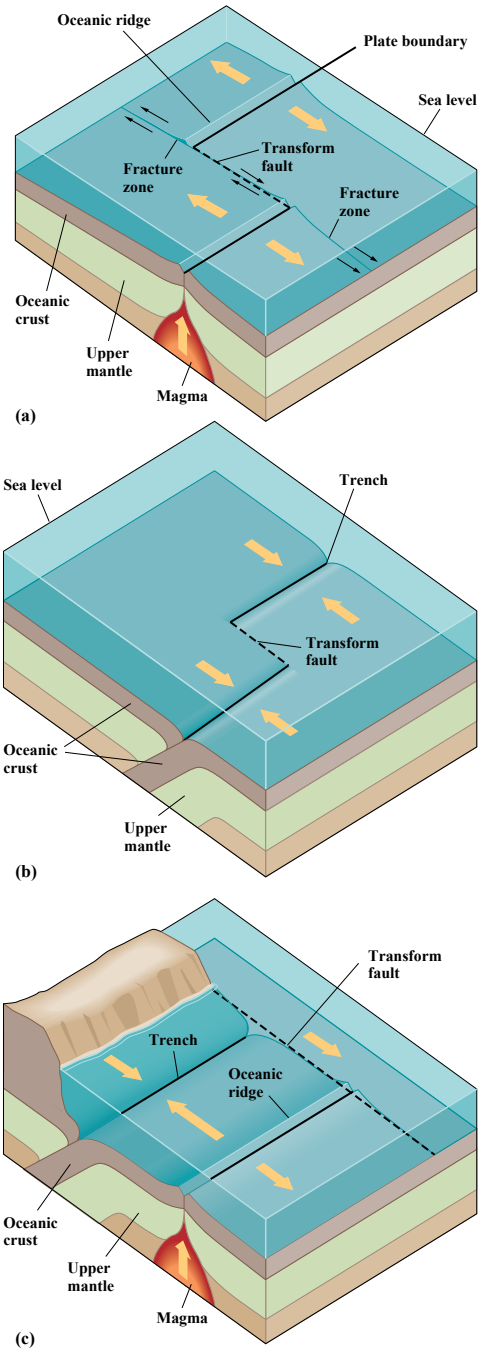

FIGURE 4-23. Adjacent plates move past each other at a transform fault. (a) Most transform faults connect two segments of an oceanic ridge. The plates slide past each other in the region between the two ridge crests. This is the transform fault. Beyond this region, the two ridge segments are locked together and form a fracture zone. A transform fault can also connect (b) two adjacent segments of a trench or two trenches, or (c) an oceanic ridge and a trench.

depressions and blanket the topography. Therefore, transform faults in subduction zones are less well defined by topography than their oceanic ridge counterparts are.

HOT SPOTS

Scattered throughout the world are locations where heat from the mantle flows outward through the crust at a much higher rate than in the surrounding crust. Some of these hot areas are within plates, and some are at plate boundaries. Most of them are beneath oceanic crust, but others are under continental crust (**Fig. 4-11**). Most hot locations occupy only a very limited area at the Earth's surface and are, therefore, called hot spots.

Many, but not all, of Earth's hot spots are known to be located where convection plumes upwell from the deep mantle (**Fig. 4-6**, **CC3**). Some of these may be plumes that extend all the way to the boundary between the core and mantle. Others may originate from a superplume in the lower mantle. Yet other hot spots, such as that at Yellowstone Park, seem to be linked to shallow convection processes in the upper mantle only. Finally, some locations where the hot area under the lithosphere is more extensive may not themselves be hot spots (although hot spots may occur within such regions). Instead, they may be locations where continents have remained for prolonged periods of time. At such locations, the continents may act as a blanket to trap the heat flowing upward through the upper mantle, and the heat may melt and thin the crust. The East African Rift may be an example. However, a superplume is believed to be located beneath the region in which the East African Rift occurs, and this rift may be the result of several interacting processes.

Most oceanic hot spots are characterized by active volcanoes that rise through the ocean floor to form islands. Examples of islands located at hot spots are Iceland and the island of Hawaii. Both have active volcanoes that generally erupt relatively quietly, without explosions, and bring copious amounts of magma to the surface (**Fig. 4-24**). The magma solidifies and steadily builds the island.

The quantity of lava produced by hot-spot volcanoes is so large that such volcanoes are the tallest topographic features on Earth. For example, the island of Hawaii rises about 5500 m from the seafloor to the ocean surface and then rises another 4205 m above the ocean surface to its highest mountain peak. Its total elevation, about 9700 m, is gained over a horizontal distance of less than 200 km and far exceeds the 8848-m elevation of Mount Everest. The enormous weight of Hawaii and the adjacent islands has depressed the Pacific Plate through **isostatic leveling** (**CC2**). As a result, the seafloor around the islands has a broad moat-like depression some 500 m deeper than the surrounding seafloor. The downward deformation of the Pacific Plate extends about 300 km from Hawaii. Surrounding the "moat" region is another broad area where the seafloor is deformed upward as the crust is compressed outward by the island rising through its center.

Until recently, it was thought that hot spots remain fixed in place with respect to the

FIGURE 4-24. (right) The 1983 eruption of Pu'u'ō'ō, the active vent of Kilauea Volcano in Hawaii. Kilauea is a hot-spot volcano that ejects small amounts of ash and large amounts of lava in a relatively smooth, non-explosive manner. This eruption of Kilauea continued for 35 years, ending in 2018.

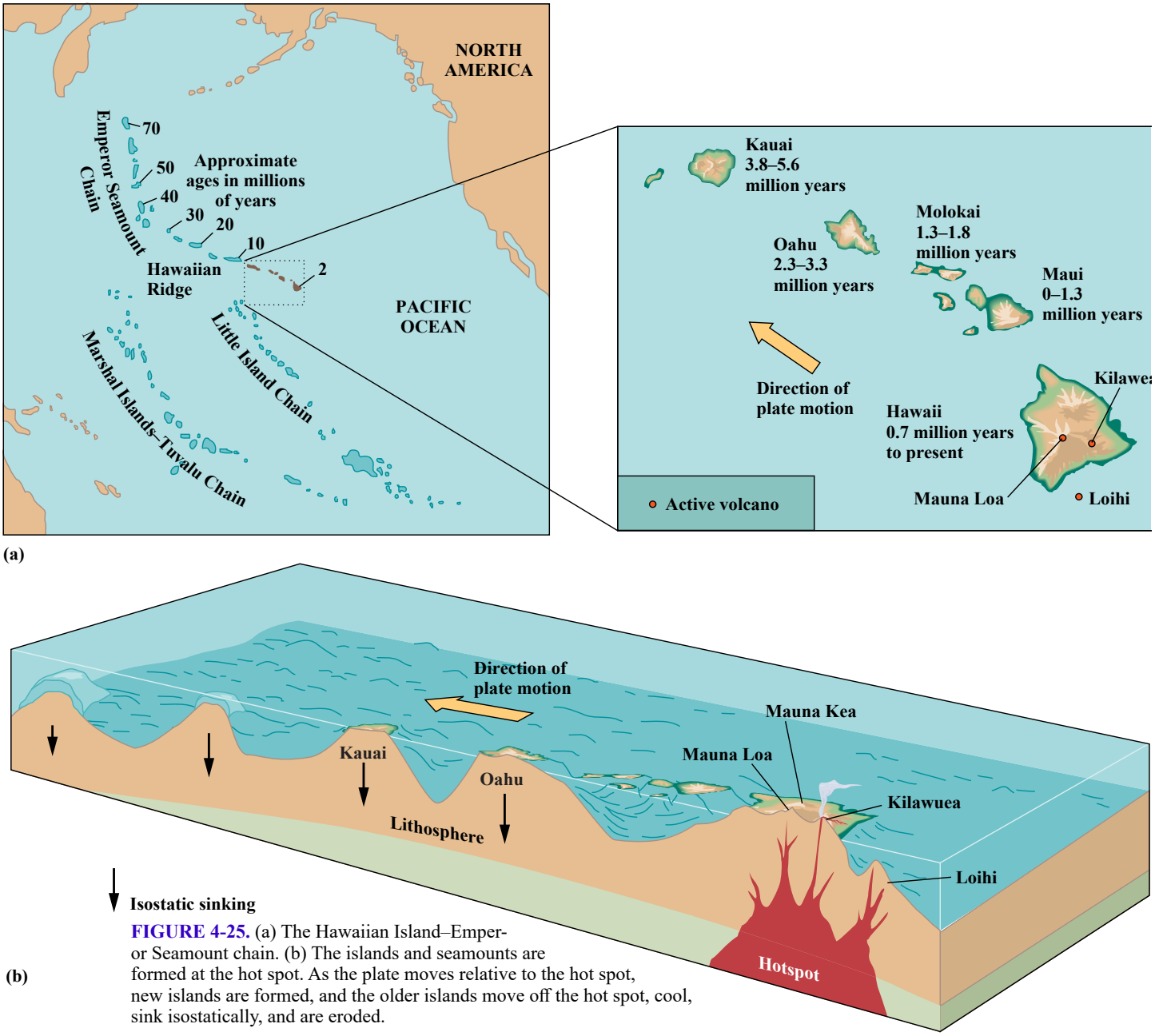

FIGURE 4-25. (a) The Hawaiian Island–Emperor Seamount chain. (b) The islands and seamounts are formed at the hot spot. As the plate moves relative to the hot spot, new islands are formed, and the older islands move off the hot spot, cool, sink isostatically, and are eroded.

Earth's rotational axis for tens or hundreds of millions of years as the lithospheric plates move over them. However, it is now known that at least some hot spots do migrate independently within the mantle, so two independent processes change the location of individual hot spots, movement of the hot spot plumes and movement of the lithospheric plates.

As a lithospheric plate moves over (or in relation to) an oceanic hot spot, the most recently formed volcanic island moves away from the hot spot. With the migration of the island from the hot spot, another segment of plate is brought over the hot spot, new volcanic vents open in it, and eventually another island may be formed. Continued plate movement takes this island away from the hot spot, and yet another island is formed, and so on. The results of that process can be seen in the seafloor topography as a trail of islands and **seamounts** (undersea cone-shaped mountains). In the Pacific Ocean, the chains of islands and seamounts that align northwest of Macdonald Seamount, Easter Island, and Hawaii are all remnants of their respective hot spots (**Fig. 4-11**).

The island and seamount trails provide a history of the rate and direction of movement of the plate relative to the hot spot. Of the Hawaiian Islands, the island of Hawaii is the youngest and has several active volcanoes, including Kilauea (**Fig. 4-24**) and Mauna Loa. About 20 km southeast of Hawaii, a 3.5 km tall seamount called Loihi has been built on the sea floor and continues to grow by volcanic action (**Fig. 4-25**). Its current peak is about 970 m below sea level, but if volcanic activity continues at its present rate, Loihi may become the next Hawaiian island 10,000 to 40,000 years from now.

Radioisotope dating (**CC7**) of the volcanic rocks in the Hawaiian Island–Emperor Seamount chain shows each island or seamount to be progressively older with distance northwest from Hawaii (**Fig. 4-25a**). For example, Oahu was formed about 2 to 3 million years ago and only has inactive, although not yet necessarily extinct, volcanoes. As hot-spot volcanic islands migrate away from the hot spot, their volcanoes become inactive, the islands cool and sink isostatically, and they are subjected to ero-

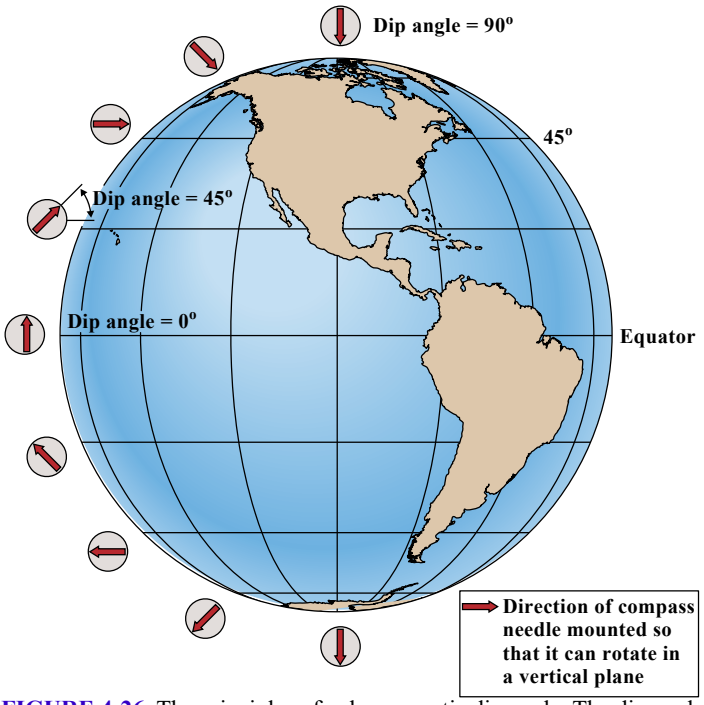

FIGURE 4-26. The principles of paleomagnetic dip angle. The dip angle is the angle between the rock's direction of magnetization and the Earth's surface (if the rock has not been tilted since being formed). Because the dip angle changes with latitude, the original dip angle of a rock's magnetization reveals the latitude at which it was formed.

sion (Fig. 4-25b).

Some hot spots lie on divergent plate boundaries where two plates are moving apart. A volcanic island formed at such a location is steadily broken apart as the plates diverge. Iceland is a good example. Each side of the island, with its cooling volcanoes, migrates away from the hot spot on its respective plate. Evidence of this process can also be seen in the ocean floor topography. For example, the Icelandic Ridge stretching between Greenland and Europe consists of sediment-buried remnants of volcanoes that occupied the Icelandic hot spot when the Atlantic Ocean was narrower. Similarly, the Walvis Ridge and Rio Grande Rise in the South Atlantic Ocean are remnants of the Tristan da Cunha hot-spot volcanoes (Fig. 4-11).

A Lesson about Science

The rate of Pacific Plate movement appears to have varied relatively little during the past 50 million years. However, a distinct change in the direction of the Hawaiian Island–Emperor Seamount chain occurs at islands formed about 50 million years ago (Fig. 4-25). Seemingly identical changes of direction are also seen in the Easter Island and Macdonald Seamount chains.

Studies of the change in direction of the Hawaiian Island–Emperor Seamount chain provide a lesson for this author and the ocean and geological science community. This story provides you, the readers of this text, with a perfect example of why critical thinking skills are so important. It also illustrates that, while the material in this book represents the most recent scientific consensus regarding what is known about our ocean world, that consensus is always evolving.

As recently as 1998 (when the first edition of this text appeared), the consensus view of the scientific community was that hot spots remained fixed in place for tens or hundreds of millions of years as the lithospheric plates moved over them. At that time, the scientific consensus was also that the distinct change in the direction of the Hawaiian Island–Emperor Seamount chain indicated an abrupt (in geological time) change in the Pacific Plate's direction of motion. Experts in the field agreed that the available data fit this conclusion. For example, the ages of the islands in the chain had been measured and did increase with distance from the hot spot. Other hot-spot trails were also found on the Pacific Plate, and they mirrored both the direction and the change in direction of the Hawaiian Island chain.

Fortunately, some scientists continued to think critically, even when the available data seemed to fit these explanations very well. These scientists were uncomfortable that nobody could explain how or why the direction of the Pacific Plate movement changed so abruptly. Thus, when the opportunity presented itself, they examined cores drilled into three seamounts of the Emperor chain that were formed before the "change in direction of motion." They analyzed these cores for many different parameters, but the key data were obtained by examination of the **paleomagnetic** signatures of the magmatic rocks that form these seamounts. These data allowed the researchers to determine the rocks' paleolatitudes (the **latitudes** at which these rocks had formed by cooling and solidification of liquid magma).

To explain how paleolatitude was measured, we have to look carefully at Earth's magnetic field characteristics. We normally hold a compass horizontally so that its needle can rotate to align north and south. However, if we turn the compass on its side and orient it toward the pole, we find that the Earth's magnetic field is not parallel to the ground (except at the magnetic equator). In the Northern Hemisphere, the compass needle points at an angle toward the ground (the **dip angle**) in the direction of the north magnetic pole (Fig. 4-26). The dip angle increases as we move toward the magnetic pole. At the magnetic pole, the needle points at a 90° angle directly into the ground. At the magnetic equator, the dip angle is zero and the needle is parallel to the ground. Because the magnetic and geographic poles are always reasonably close together, the dip angle of the compass needle is a measure of latitude. Larger angles correspond to higher latitudes.

Magnetic materials in magma orient both horizontally and vertically to align with the Earth's magnetic field when the magma solidifies. Therefore, rocks carry paleomagnetic information about both their horizontal orientation with respect to the pole at the time they were formed, and the latitude at which they were formed, the paleolatitude.

What the measured paleolatitudes of the three Emperor chain seamounts revealed was that these islands had not been formed at the same latitude that Hawaii now occupies. Instead, each older seamount had been formed progressively farther north. How could this be if hot spots are fixed in place? Of course, it cannot. Most scientists had accepted the hypothesis that hot spots remain fixed in place relative to the Earth's axis of rotation. That hypothesis now appears to be wrong. This hot spot must have moved south during a period about 50 to 80 million years ago, when the Emperor seamounts were formed. The lesson here is that, when a new hypothesis is proposed, there are almost always a few pieces of data missing, or data that do not exactly fit the "facts" of the hypothesis. These minor inconsistencies are the clues to how a hypothesis might be wrong.

Where does this new evidence leave our understanding of hot spots and how they fit into the jigsaw puzzle of plate tectonics? First, it raises many more questions. For example, why are there almost identical changes in direction in other Pacific hot-spot trails? Did all the Pacific hot spots move in the same direction at the same time? Scientists must answer these questions, and many more, if we are to better understand plate tectonic processes.

PLATE INTERIORS

Oceanic crust is formed and destroyed by the tectonic processes that occur at divergent and convergent plate boundaries and hot spots. Its topography is modified by various processes over the tens or hundreds of millions of years between its creation and destruction. The edges of the continents, which are formed at divergent plate boundaries, are modified by a variety of processes between the time they are formed and the time they enter convergent plate boundaries.

Oceanic Crust

As it moves away from an oceanic ridge, the lithosphere cools, becomes denser, and sinks steadily deeper into the asthenosphere. Therefore, the seafloor becomes lower with increasing distance from the oceanic ridge. As the cooling crust sinks, it is progressively buried by a continuous slow "rain" of solid particles through the water column that accumulate as sediments on the seafloor (**Figs. 4-20, 4-27**). Because sediments tend to accumulate faster in topographic lows, the original topography of the rugged oceanic crust that is formed at oceanic ridges is progressively buried and smoothed as the crust moves away from the oceanic ridge. The effect is very similar to that of a snowfall. If left undisturbed, a few centimeters of snow obscures features, such as street curbs and potholes, and softens larger features by mounding around them. As more snow accumulates, larger features are buried, and even cars in a parking lot may be difficult to find. Similarly, the lower topography of the oceanic crust is completely obscured after it has traveled a few hundred kilometers away from the ridge, though the oceanic crust takes millions of years to move such distances. The higher topography of the oceanic crust survives the sedimentation as rounded hills or mountains rising above the surrounding flatter areas.

The largest topographic features of oceanic hot spots and convergent boundaries where both plates have oceanic crust (e.g. magmatic arcs) are commonly cone-shaped volcanoes. Their conical form is preserved and even enhanced by sediment accumulation. Therefore, much of the deep-sea floor is characterized by generally cone-shaped abyssal hills and mountains (called seamounts), even in regions far from the oceanic ridges or hot spots (**Fig. 3-4**).

Some seamounts have flat tops and are called "tablemounts" or **guyots**. Volcanic mountain cones that have sufficient elevation when first formed at the oceanic ridge or at hot spots emerge above sea level as islands. The island tops are eroded by wind, water, and waves much faster than the volcano can cool and sink isostatically. Thus, before such volcanoes sink isostatically below the surface, their tops are totally eroded away. The eroded flat tops are preserved once the volcanoes are completely submerged because erosion is extremely slow under the ocean surface, far away from winds and surface waves. The Hawaiian Island–Emperor Seamount chain exemplifies the various stages of this process (**Fig. 4-25**).

Some seamounts form the submerged base of nearly circular coral reefs called atolls (**Fig. 4-28**). Reef-building corals grow only in shallow water (less than several hundred meters), and most species inhabit the warm tropical oceans (**Chap. 15**). In such regions, a coral reef may become established around islands that are formed when the tops of oceanic ridge or hot-spot volcanoes extend to or above sea level (**Fig. 4-29**). The coral reef

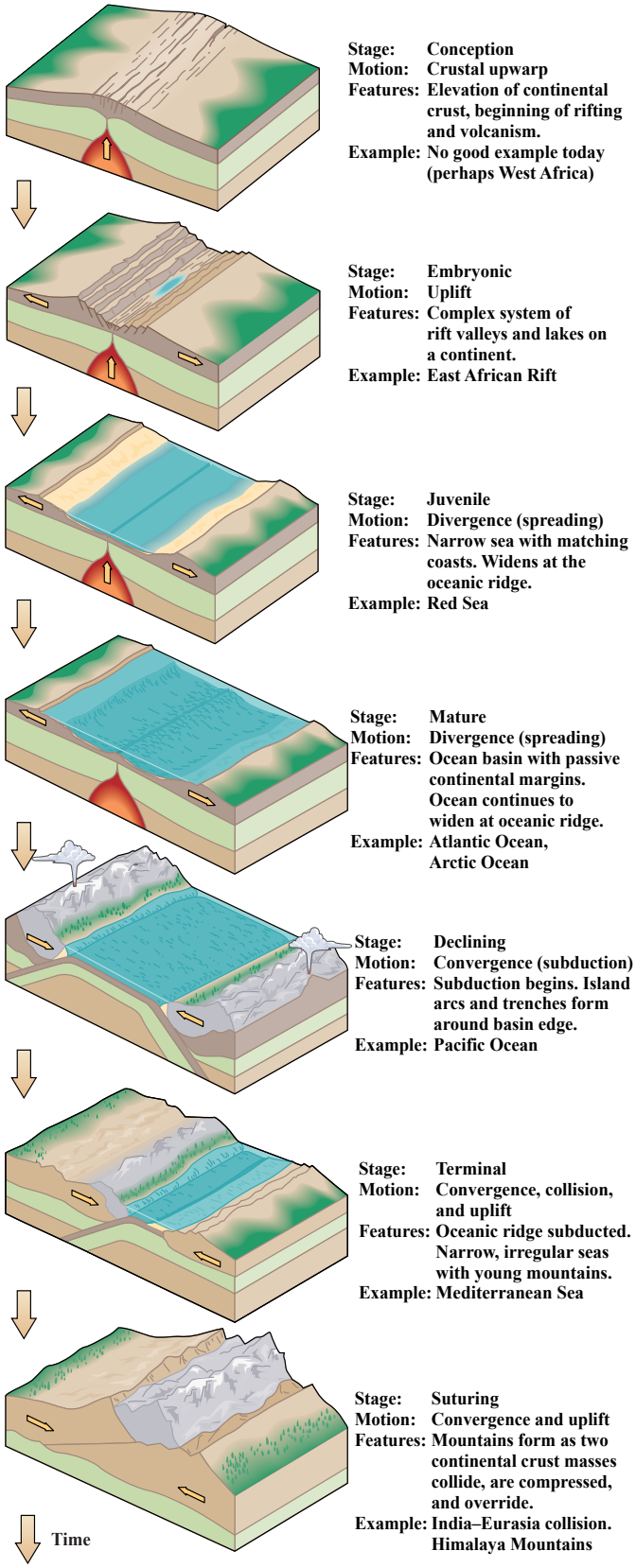

FIGURE 4-27. The history of an ocean from its creation at an oceanic ridge to its disappearance at a continental collision.

FIGURE 4-28. Satellite image of a coral atoll with its shallow interior lagoon (Hull Island, Kiribati).

continues to build upward from the flanks of the volcano as the volcano sinks isostatically. If the upward growth rate of the reef is fast enough to match the sinking rate of the volcano, the top of the live coral remains in sufficiently shallow water to continue growing.

The coral first forms a **fringing reef** adjacent to the island (**Fig. 4-29a**). As the volcano continues to sink, this reef continues to grow upward and becomes a barrier reef separated from the island by a lagoon (**Figs. 4-29b, 4-30**). Eventually, the volcano completely sinks and leaves only an atoll (**Figs. 4-28, 4-29c**). On the historic Beagle voyage of 1831 to 1836, Charles Darwin visited Keeling Atoll and several reef-fringed islands of the South Pacific. Even though Darwin had no knowledge of plate tectonics, he proposed an explanation for the formation of atolls that is very similar to our understanding of that process today.

Continental Edges

We have seen how the processes at convergent, divergent, and transform fault plate boundaries form and shape the edges of continents. However, fewer than half of the coasts (or margins) of today's continents lie at plate edges. Most are located in the middle of lithospheric plates. Most of these coastlines were formed initially at the rift zones created as Pangaea broke apart. Because there are few earthquakes or volcanoes at such **continental margins**, they are called **passive margins**. The Atlantic coasts of North America and western Europe are examples.

Passive Margins

Development of a passive margin begins as the edge of a new continent is formed at a mature continental rift valley (**Fig. 4-31**). The new edge is isostatically elevated because of heating in the rift zone. Consequently, rivers drain away from the edge toward the interior of the continent. Shallow seas form in the rift as blocks of the continental crust edge slide down into the rift zone. Because few rivers carry sediment into the seas, the **turbidity** of the water in these seas is low. Consequently, the light needed for **photosynthesis** (**CC14**) penetrates deep into the clear waters, and **primary productivity** (growth of marine organisms that create organic matter from carbon dioxide and an energy source; **Chap. 12**) is high. High productivity leads to large quantities of organic matter that may accumulate in the sediment of the new marginal seas. Because the seas are shallow and the rift valley is narrow, they may be periodically isolated from exchange with the large ocean basins. Under these conditions, thick salt deposits may be formed as seawater evaporates and its dissolved salts precipitate (**Chap. 6**).

As the passive margin moves away from the divergent plate boundary, which by that time has developed an oceanic ridge, both continental crust and oceanic crust cool and subside isostatically. Eventually, the edge of the continent subsides sufficiently that the direction of the slope of the landmass is reversed. Then the rivers that flowed away from the margin during its early history instead flow into the ocean at the margin. This **runoff** brings large quantities of sediment from the land, which increases turbidity and siltation in the coastal ocean, thus reducing light penetration and primary productivity. Both the marginal seas and their deposits of organically rich sediment eventually are buried deeply as they subside farther below sea level. When organic sediments are sufficiently deep to be heated and subjected to high pressure, the organic matter may be converted into **hydrocarbon**

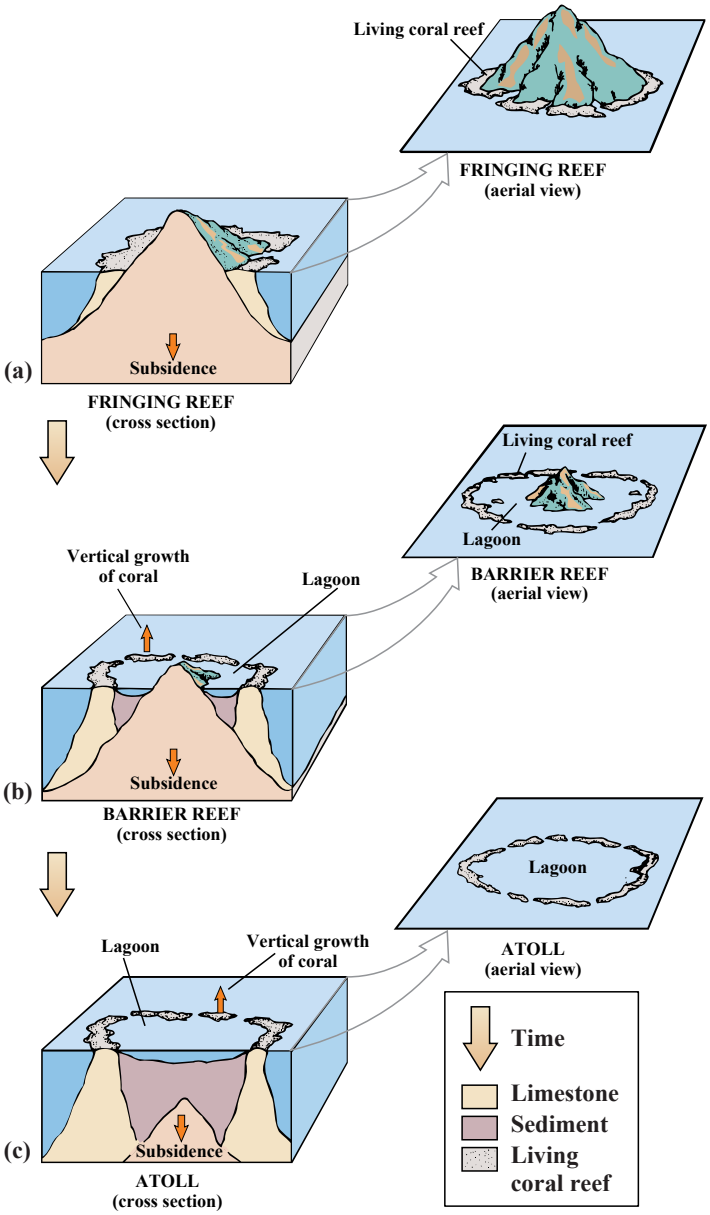

FIGURE 4-29. Coral reefs form in shallow water in the tropics and subtropics. Reefs are well developed around volcanic islands created at hot spots because there is little runoff of freshwater and sediment, both of which inhibit healthy growth of coral reefs. Coral reefs evolve as volcanic islands are formed and then sink isostatically as they move off the hot spot and cool. (a) Fringing reefs are formed around the perimeter of rising or static volcanic islands. (b) When the island sinks, a barrier reef is formed as the fringing reef grows upward. (c) Eventually, the island sinks completely beneath the surface, leaving an atoll where the barrier reef continues to grow upward.

FIGURE 4-30. This volcanic island, Bora Bora in the Leeward group of the Society Islands of French Polynesia, is surrounded by a lagoon and a well-developed barrier reef.

compounds, such as those found in oil and gas. If the overlying rocks are permeable, the oil and gas migrate toward the surface. Where the overlying rocks are not permeable, they trap the oil and gas, forming reservoirs. Reservoirs formed at passive margins provide most of the world's oil and gas.

The characteristic feature of a passive margin is a **coastal plain**, which may have ancient, highly eroded hills or mountain chains inland from it. In addition, the coastal region is generally characterized by **salt marshes** and many shallow estuaries. Offshore, the continental shelf is wide and covered by thick layers of sediment. These features have been modified in many areas by changes both in sea level and in the distribution of glaciers that, in turn, were caused by climate changes.

The Fate of Passive Margins

The depth and width of the continental shelf at passive margins can be influenced by isostatic changes caused by the heating of relatively immobile continental blocks (CC2). For example, the continental shelf of much of the west (Atlantic) coast of Africa is narrower than those of many other passive margins. The reason is that the entire continent has been lifted by isostatic leveling in response to the heat accumulation that is causing the East African Rift to form. Africa's Atlantic coast may eventually become a new subduction zone if the Atlantic Ocean crust near Africa cools sufficiently and if the East African Rift continues to expand. Another possibility is that the rifting in East Africa will simply stop. Many factors will influence the outcome, including how much heat is available in the mantle to drive the East African

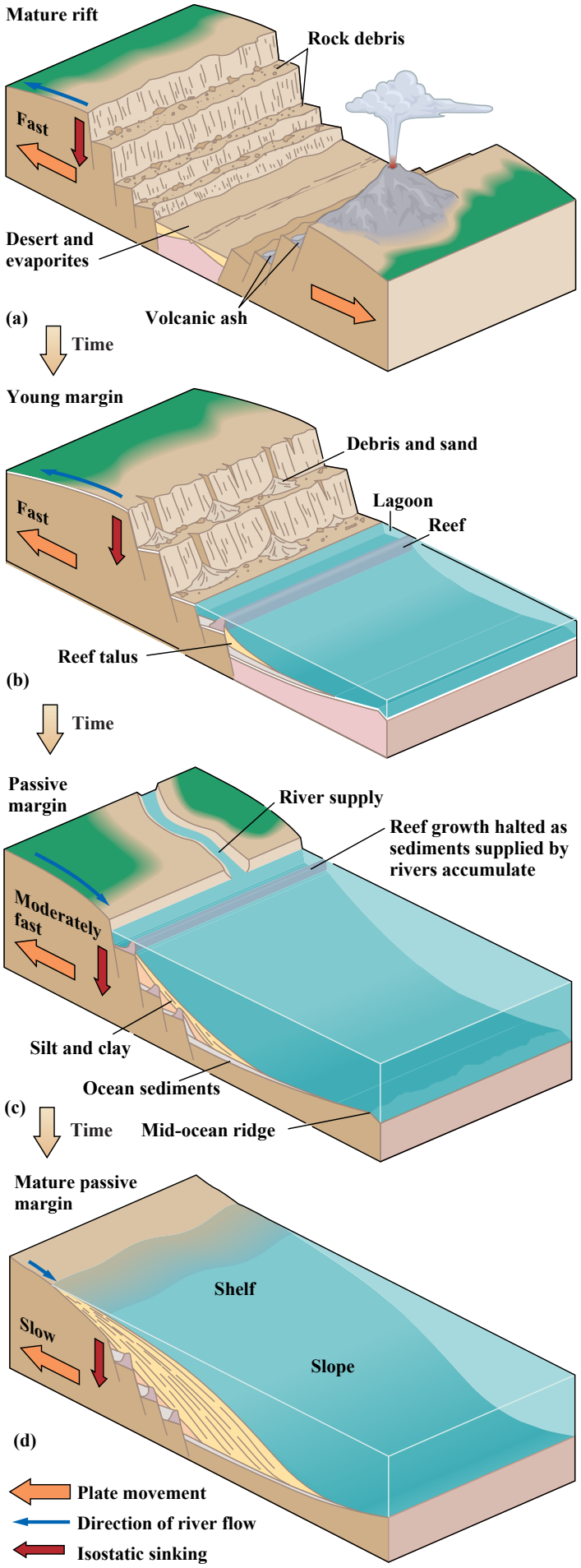

FIGURE 4-31. A passive margin develops after the initial stages of the creation of a new ocean. (a) A rift forms at a continental margin. Layers of volcanic ash, desert sands, and evaporites form as lakes develop and dry up in the rift valley. (b) As the new ocean forms, reefs and high-productivity lagoons form along the ocean edge because rivers flow inland and turbidity is low in such young oceans. (c) As the margin moves away from the new oceanic ridge, it cools and subsides. Rivers start to flow toward the new ocean, depositing lithogenous sediments in the previously productive lagoons and reefs. Productivity in the coastal waters, especially coral reef growth, is suppressed by high turbidity caused by the river inputs of suspended sediment. (d) Sediment layers build up on the passive margin to form a broad sediment-covered continental shelf. The buried lagoon and reef sediments often form oil reservoirs.

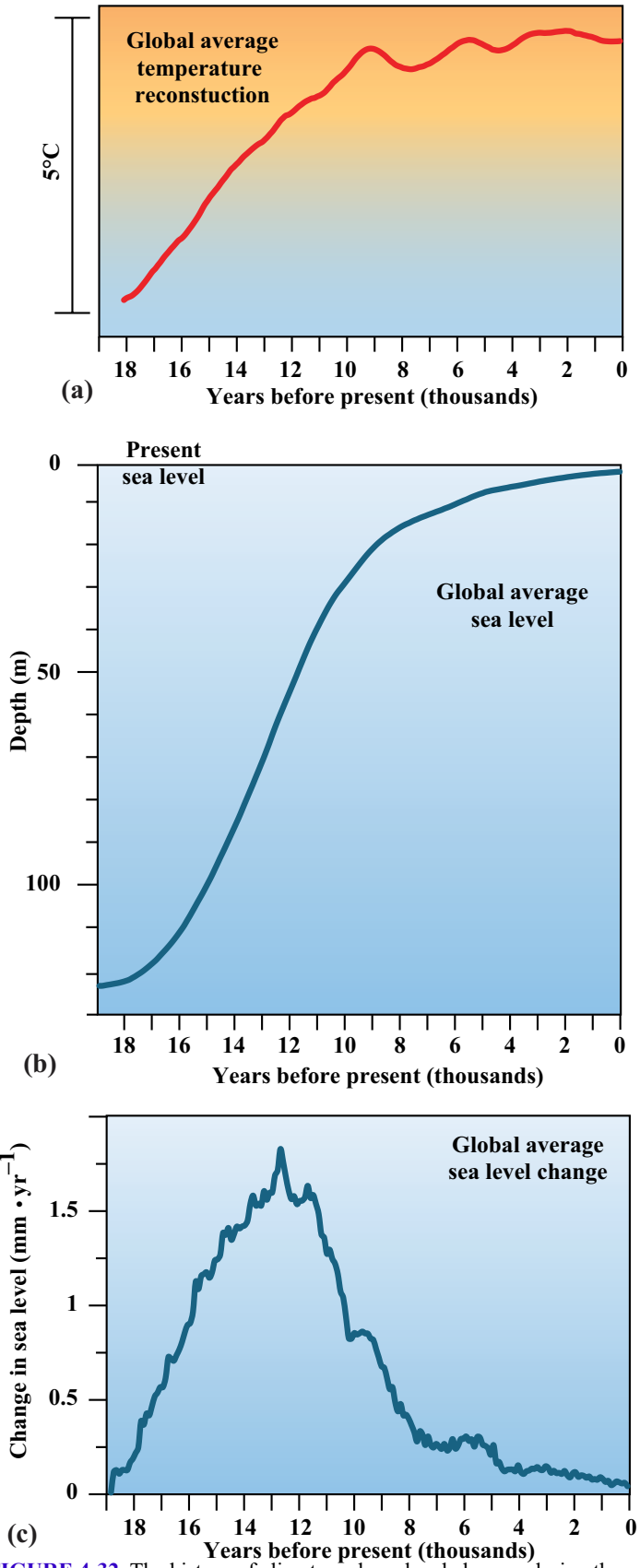

FIGURE 4-32. The history of climate and sea-level changes during the past 18,000 years. (a) The global average surface air temperature was about 7–8°C lower than today until about 18,000 years ago and rose close to its present value by about 8000 years ago. (b) In response to the global temperature increase, sea level rose rapidly (about 120 m in total) as global climate warmed from about 18,000 to about 8000 years ago. More recently, sea level continued to rise slowly as global temperatures remained relatively stable. (c) The rate of sea level change increased to nearly 2 mm•yr^{-1} about 12,000 years ago and has decreased since then.

rifting process and to continue spreading at the Mid-Atlantic Ridge, how cold and dense the Atlantic Ocean crust is near West Africa, and the driving forces on other plates.

Whatever happens in Africa, the probable ultimate fate of some passive margins is that they will become subduction zones. This will be the outcome if the oceanic crust at the passive margin cools sufficiently to subduct into the mantle, causing the plate movements to change and the ocean to begin to close. Eventually, the ocean could be destroyed as two continents meet at a collision margin

SEA-LEVEL CHANGE AND CLIMATE

Although tectonic processes and sedimentation are the principal agents that shape seafloor topography, changes in climate also affect certain features, particularly the continental shelves and coasts. Substantial climate changes have been observed during recorded human history. For example, much of North Africa and the Middle East, which is now a region of desert and near-desert, was a fertile region with ample rainfall during Greek and Roman times. However, such short-term changes are small in comparison to changes that occur during the hundreds of millions of years of a tectonic spreading cycle.

Climate Cycles

Over the past 1000 years, the Earth's global average surface air temperature has varied by about 1.5°C. During the past 10,000 years, it has varied within a range of about 2°C to 3°C (**Fig. 4-32**). However, in the immediately preceding 2 to 3 million-year period, the Earth's temperatures were as much as 10°C below what they are today. These colder periods are called **ice ages** or glacial periods. During the most recent ice age, the Earth's climate alternated between periods of glacial maxima (when temperatures were about 10°C below those of today) and interglacial periods (when temperatures were close to those of today). During glacial maxima, the polar ice sheets extended to much lower latitudes than they do now. Several earlier major ice ages each lasted about 50 million years. Longer periods with warmer climates have occurred between ice ages. The most recent ice age may not yet be over, and the relatively warm climate of the past 10,000 years may simply represent a geologically short interglacial period.

Climate varies on timescales ranging from a few years to tens or hundreds of millions of years. Variations over the past 18,000 years are shown in **Figure 4-32**. These shorter timescale variations are superimposed on longer-term variations. Understanding the causes and consequences of the historical variations is important. Such knowledge may provide the key to understanding the effects and consequences of the extremely rapid (on geological timescales) global warming that has been predicted to occur and that appears to already have begun as a result of human enhancement of the atmospheric **greenhouse effect**.

Eustatic Sea-Level Change

When the Earth's climate cools for a long time, ocean water cools and contracts and polar ice sheets expand as water is transferred from oceans to land in the form of glaciers and snow. Both of these processes cause the sea level to drop. Conversely, warm climate periods tend to heat and expand ocean water, and melting continental ice returns water to the oceans, raising the sea level. Such changes of sea level, called eustatic changes, occur synchronously throughout the world, although not all locations experience exactly the same amount of rise or fall. In contrast, sea-level

changes caused by isostatic movements of an individual continent affect only that continent (CC2).

During the initial breakup of Pangaea, the climate was relatively warm, and it remained so until about 10 to 15 million years ago. During the warmest part of this period, about 75 million years ago, sea level was considerably higher than it is today, and as much as 40% of the Earth's present land area was below sea level. For example, a shallow sea extended from the Gulf of Mexico far north into Canada and covered what is now the land between the Rocky Mountains and the Appalachian Mountains. In contrast, at the peak of the most recent glacial period, about 20,000 years ago, sea level was at least 100 m below its present level, exposing much more land area. Sea level has risen and fallen by various amounts many times as Pangaea has broken apart in the present spreading cycle, and coastlines have migrated back and forth accordingly.

Sea-Level Change and Continental Margin Topography

Erosion by rivers carves out valleys, and the rivers transport the eroded sediment downstream, where it is deposited in lower-lying areas or in the coastal oceans. Erosion by waves and winds at coastlines (Chap. 11) also tends to reduce topography and deposit the eroded sediment in the shallow waters of the continental shelves. In contrast, at water depths of more than a few meters, erosional forces in the oceans are generally reduced and sediment accumulation is dramatically reduced, except in proximity to rivers that transport massive sediment loads to the ocean (Chap. 6).

As a result of sea-level oscillations, the area between the edge of the continental shelf and an elevation several tens of meters above the present sea level has been subjected to alternating cycles of wind, river, and wave erosion at some times and sediment deposition at others. These processes have substantially modified the topography along the coasts. The effect is most apparent at passive margins. On most such margins, there is evidence of sea incursion and erosion throughout the area between the continental shelf edge and areas far inland from the current coastline. This evidence includes buried deposits that contain freshwater and shallow-water marine organisms and flat or low-relief topography.

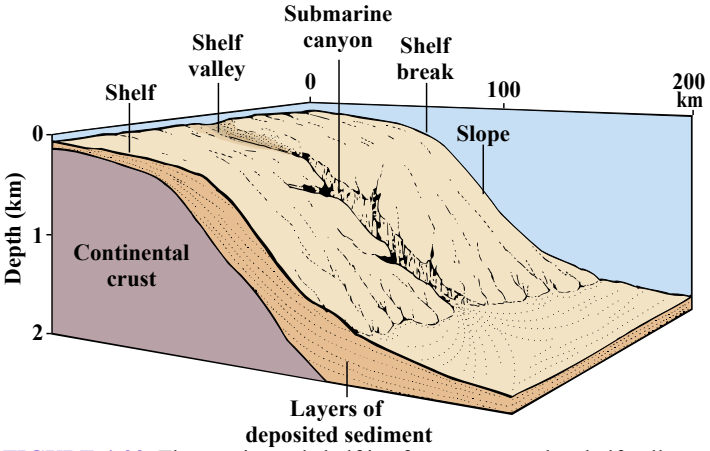

FIGURE 4-33. The continental shelf is often cut across by shelf valleys, some of which connect to steep-sided submarine canyons that cut across the continental slope.

The continental shelf, which has been progressively covered by the rising sea during the past 19,000 years, is cut across by numerous **shelf valleys**. Most of the valleys were carved out by rivers during the last ice age, when sea level was lower. Many **submarine canyons** are extensions of shelf valleys (Fig. 4-33).

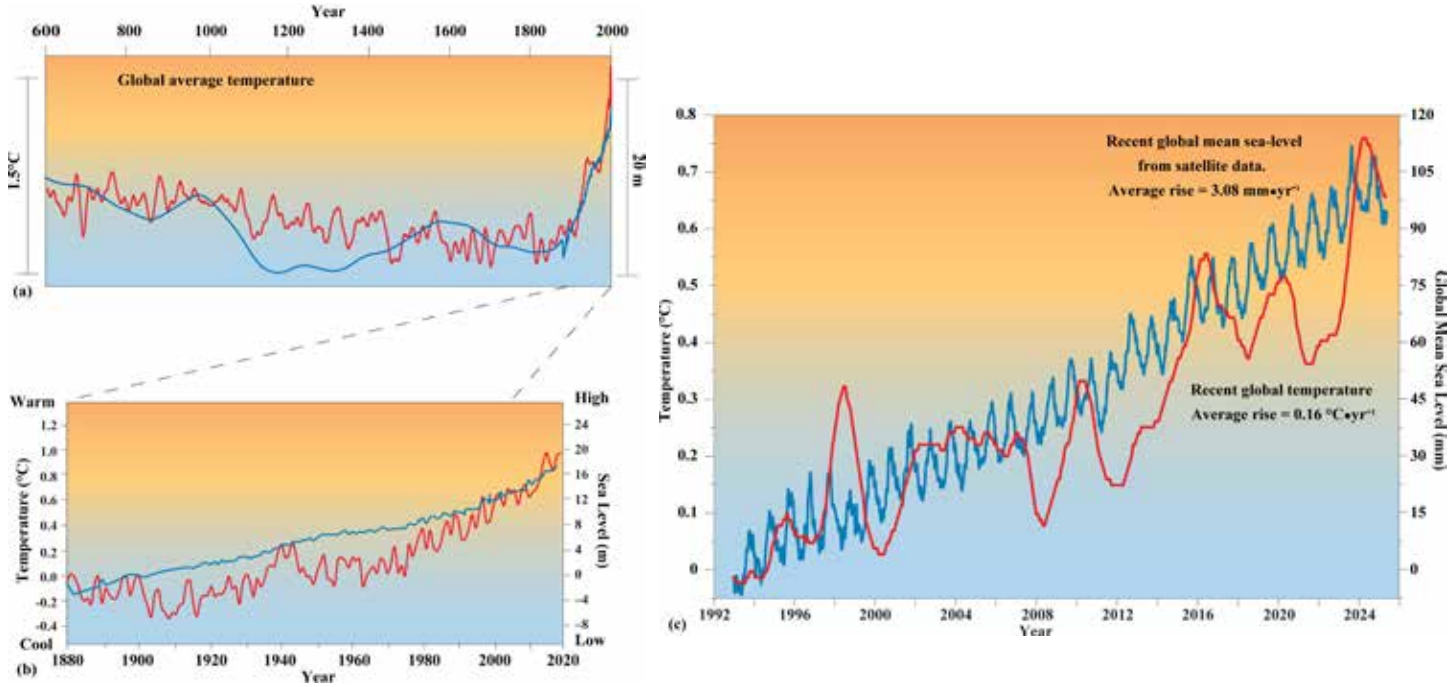

FIGURE 4-34. Global average surface air temperature and sea level variations within the past 1400 years. (a) Since 600 CE, global average surface air temperature has varied by about 1.5°C, but is currently rising at about 1°C per century. Sea level has varied by about 20 m and appears to follow the trend of temperature changes. (b) Since about 1920, air temperature has started to rise more quickly, as has sea level. However, there is much short-term variability, and the apparent upward trend may not be sustained. (c) Since 1993, instrumental records of temperature show large variations in temperature, but a general increasing trend of 0.16°C•yr^{-1}. During the same period, highly-accurate satellite measurements show that sea level rise may be accelerating and is now estimated to be about 3 mm•yr^{-1}. It is not known how far temperature and sea level will rise in the future. It is known that the temperature increase and sea level rise are due to a combination of natural change and the enhanced greenhouse effect but it is not clear what the relative contributions of these two factors has been. This raises the question as to whether the rate of sea level rise could accelerate substantially if the current rising temperature trend is sustained.

Isostatic Sea-Level Change

In addition to eustatic processes, isostatic processes induced by changing climate can affect coastal topography. During an ice age, massive ice sheets accumulate over many parts of the continental crust. The weight of the ice forces the continental crust to sink lower into the asthenosphere (**CC2**). This process also depresses (lowers the level of) the continental shelves. For example, Antarctica's continental shelf is currently depressed about 400 m lower than most other shelves by the weight of its ice sheets. At the end of an ice age, when the ice melts, a depressed section of a plate slowly rises until it again reaches isostatic equilibrium. However, isostatic leveling is much slower than eustatic sea-level change. Many areas of continental crust, including the coasts of Scandinavia and the northeastern U.S. and Canada, are still rising in response to the melting of glaciers that occurred several thousand years ago.

Glaciers shape topography in other ways as well. They often cut narrow, steep-sided valleys. Many valleys left after glaciers melted have been submerged by rising sea level to become deep, narrow arms of the sea known as **fjords** (**Chap. 13**).

Sea Level and the Greenhouse Effect

Although oscillations of sea level are normal occurrences in geological time, human civilization has emerged during a long period of relatively stable sea level (**Fig. 4-34**).

If the predicted greenhouse warming of the planet by as much as 2°C during the first half of this century does indeed occur, the higher average global temperature will cause the ocean water to continue to warm and expand, and the current polar ice sheets to continue to melt. As a result, sea level could rise by several meters or more. Warming and melting of ice sheets are slow processes that will continue possibly for centuries even if we were to completely stop adding to the concentration of carbon dioxide in the atmosphere. However, the most recent predictions based on the best available scientific information are that sea level will rise by about 1 m, and likely more, by the end of this century. A sea level rise of only a few tens of centimeters would inundate vast areas, including major coastal cities. Therefore, the history of coastal modification during periods of sea-level rise is of more than academic interest.

PRESENT-DAY OCEANS

Having learned about processes that create, shape, and destroy ocean floor topography, we can look with greater understanding at the present-day oceans, which are all connected. The Atlantic Ocean connects with both the Indian and Pacific Oceans near Antarctica. The Arctic Ocean connects with the Pacific Ocean only where shallow water covers the continental shelf in the Bering Strait between Alaska and Siberia. In contrast, the Arctic Ocean is connected with the Atlantic Ocean by the deeper and much wider passage through the Nansen Fracture Zone between Greenland and Spitsbergen, Norway (**Fig. 4-18**). The somewhat deeper connection between the Arctic and Atlantic Oceans allows somewhat restricted water transfer between the two basins (**Chap. 8**).

Pacific Ocean

The Pacific Ocean is the world's largest and, on average, deepest ocean. It is almost completely surrounded by narrow continental shelves and deep trenches of subduction zones. It has many islands, including volcanic islands and atolls formed at hot spots, islands in magmatic and sedimentary arcs, and islands that appear to be small pieces of previous continents.

Aside from hot spots and limited areas of new spreading, notably the East Pacific Rise, the Pacific Ocean crust is generally old, cold, and dense. Hence, it floats low in the asthenosphere. However, the oldest oceanic crust found in the Pacific Ocean is not substantially older than the oldest oceanic crust in other oceans. This is because all of the Pacific Ocean's older crust has been subducted. Passive margins of continents that once surrounded the Pacific developed into subduction zones long ago. Oceanic crust that once was present near the passive margins also has been subducted. The sediment that once covered the subducted crust was either subducted or compacted into rock and added to island arcs or continents as exotic terrane.

When the Pacific Ocean margins changed from passive margins to subduction zones as the ocean stopped expanding hundreds of millions of years ago, a ring of coastal mountains formed on the surrounding continents. The mountains prevented most rivers on these continents from draining into the Pacific. In addition, trenches in which river-borne sediments accumulated were formed around much of the Pacific. In other areas, magmatic arcs formed, creating marginal seas where sediments discharged by rivers were trapped. All oceanic crust now found in the Pacific Ocean, other than crust formed in marginal basins that are not part of the Pacific Plate, was created after these mountain chains, trenches, and magmatic arcs were formed. Hence, since its formation, the oceanic crust of the Pacific Plate has remained remote from any major rivers and the large amounts of sediment that they can contribute (**Chap. 6**), and the **sedimentation rate** on this crust has been slow for hundreds of millions of years. Because it is not significantly older than the crust in other oceans and because the sedimentation rate is low, the Pacific Ocean floor has a relatively thin cover of sediment. As a result, the abyssal plains are not as flat as parts of the Atlantic Ocean's abyssal plains.

Atlantic Ocean

The Atlantic is a long, narrow ocean that is often considered to include the Arctic Ocean because the Mid-Atlantic Ridge stretches essentially from the North Pole to Antarctica. The Atlantic Ocean is expanding in an east–west direction. An oceanic ridge, the Mid-Atlantic Ridge, runs down its center. Most of the Atlantic coasts of Europe, Africa, and North and South America are passive margins. The Atlantic's oldest oceanic crust is at two subduction zones: the Puerto Rico–Cayman Trench at the entrance to the Caribbean Sea, and the South Sandwich Trench located northeast of the Antarctic Peninsula.

The Atlantic Ocean has fewer island arcs and hot spots than the Pacific has. Greenland, the world's largest island, is part of the North American continent and is joined to Canada by the submerged continental shelf of Baffin Bay. Continental shelves are broader in the Atlantic Ocean than in the Pacific Ocean, and they constitute a larger percentage of the total area of the Atlantic than of the Pacific. The Atlantic also has several shallow marginal seas, including the Baltic, Mediterranean, and Caribbean Seas and the Gulf of Mexico. Thick salt deposits and oil and gas reservoirs typical of developing passive margins are present in many locations around the edges of the Atlantic Ocean. Such deposits formed early in the ocean's history, during periods when the new ocean and its marginal seas were isolated from the rest of the world ocean.

The Atlantic Ocean floor has fewer seamounts and a smoother abyssal plain than the Pacific Ocean floor, in part because of the relative dearth of hot-spot volcanoes in the Atlantic. In addition, topography is buried by very large quantities of sediment deposits that are found on the Atlantic Ocean crust, except on the newly formed oceanic ridge mountains. The large quantities of sediments are derived from the great quantities of sediment that rivers have transported into the Atlantic, as the passive margins on both sides of the ocean have eroded. The Atlantic continues to receive freshwater runoff and some sediment from vast drainage areas of Europe, Africa, and the Americas, particularly in equatorial regions. Two large rivers, the Amazon and the Congo, empty into the equatorial Atlantic. They contribute about one-fourth of the total worldwide river flow to the oceans.

Indian Ocean

The Indian Ocean is the youngest of the three major ocean basins; it was formed only during the past 125 million years by the breakup of Gondwanaland. The plate tectonic features and history of the Indian Ocean are complex. For example, it is not yet known how and why the prominent Ninety East Ridge (**Fig. 4-11, Fig. 3-4**) that divides the Indo-Australian Plate was formed.

The northern part of the Indian Ocean is dominated by the collision plate boundary between India and Eurasia. The newly formed Himalaya mountains are readily eroded, and large quantities of sediment are transported into the Arabian Sea and the Bay of Bengal by many rivers. Those rivers include three that are among the world's largest: the Indus, which empties into the Arabian Sea; and the Ganges and Brahmaputra, both of which empty into the Bay of Bengal. The enormous quantities of sediment flowing from India and the Himalayas have accumulated to form massive **abyssal fans** and extensive, relatively shallow abyssal plains in large areas of the northern Indian Ocean.

To the east of the India–Eurasia collision, along the northeastern edge of the Indian Ocean, is the very active Indonesian subduction zone. Its trenches and island arcs extend between mainland Asia and Australia. To the west of the India–Eurasia collision, the northern Red Sea is a rift zone that becomes an oceanic ridge system at its southern end. This oceanic ridge extends southward into the central Indian Ocean, where it divides. One of the ridges that originates at that point extends southwest around Africa, and the other extends southeast around Australia. The Indian Ocean is opening in a complex manner as the African, Antarctic, and Indo-Australian plates move apart.

Passive margins are present along most of West Africa, most of Australia, and the coast of India. There are few islands in the Indian Ocean other than the many islands that form the Indonesian arc, where subduction is occurring along an oceanic convergent plate boundary. Madagascar, which is now part of Africa, appears to be a fragment of Pangaea that broke away from India when India began to move rapidly northward toward Asia, long after the initial breakup of Pangaea itself.

Marginal Seas

Several arms of the major oceans are partially isolated from the major ocean basins by surrounding landmasses. Such marginal seas are of four types:

One type consists of shallow submerged areas of continental crust. Examples include the Baltic Sea, the North Sea, Baffin Bay, and Hudson Bay.

A second type is formed in back-arc basins behind subduction zones, which often contain deep areas where oceanic crust is present. The marginal sea is separated from the open oceans by islands of the magmatic arc and a submerged ridge that connects the islands. Many such seas contain thick sediment deposits derived primarily from erosion of the newly formed islands of the arc. Examples include the Java Sea behind the Indonesian Arc, the South China Sea behind the Philippines, and the Aleutian Basin in the Bering Sea behind the Aleutian Island chain.

The third type of marginal sea is a narrow remnant of an old closing ocean. The Mediterranean Sea, which lies between the converging African and Eurasian Plates, is a good example. Such seas may have thick sediments.

Finally, some marginal seas are the long and narrow arms formed as a continental divergent plate boundary develops into a new ocean. The primary example is the Red Sea.

CHAPTER SUMMARY

The Earth and Plate Tectonics.

The Earth consists of a solid inner core, liquid outer core, plastic mantle, and solid overlying lithosphere. The mantle, especially the upper mantle or asthenosphere, is close to its melting point and can flow like a fluid, but very slowly. The thin lithosphere consists of continental or oceanic crust overlying a layer of solidified mantle material and is separated into plates that float on the asthenosphere. About 225 million years ago, all the continents were joined. Since then they have been separated by plate tectonic movements.

Plate Boundary Processes.

Lithospheric plates may pull away from (diverge), collide with (converge), or slide past each other. Oceanic crust is created at divergent plate boundaries and destroyed at convergent plate boundaries. Divergent plate boundaries are the oceanic ridges and areas where continents are being pulled apart. Oceanic ridges are undersea mountain chains with many active volcanoes, and they are offset at transform faults.

Convergent plate boundaries are downwelling zones where old oceanic crust is subducted. Subduction zones at the edges of continents are characterized by an offshore trench and coastal mountains formed by compression of the continental crust plate and accumulation of sediment scraped off the subducting oceanic crust plate. Subducted and heated crust melts and magma rises to form volcanoes on the continental crust plate. Subduction zones at which oceanic crust is at the edge of both plates are characterized by a trench and a magmatic arc (and sometimes a separate sedimentary arc) of islands on the nonsubducting plate. A high subduction rate may stretch the nonsubducting plate forming a back-arc basin. A collision where two continents meet at a convergent plate boundary is characterized by mountain chains created by compression of the continental crusts of the two colliding plates.

Hot Spots.

Hot spots cause persistent volcanic activity. Some are situated over zones where upwelling convection extends throughout the mantle. Lithospheric plates move independently of any movements of most hot spots. As the lithospheric plate, hot spot, or both move with respect to each other, hot-spot trails of islands and seamounts are formed.

Plate Interiors.

As new oceanic crust moves away from a divergent plate

boundary, it cools, sinks isostatically, and is buried by sediment. Edges of continents that are not at plate boundaries are known as passive margins and are characterized by a flat coastal plain, shallow estuaries and swamps, and a wide, heavily sediment-covered continental shelf.

Sea-Level Change and Climate.

The Earth's climate is naturally variable. When the average surface temperature changes, eustatic changes of sea level occur globally. When the Earth warms, sea level rises as ocean water expands thermally, and vice versa. At the Earth's warmest temperatures, the oceans covered as much as 40% of the present land surface area. At its lowest temperatures, sea level was at least 100 m lower than it is today, and most of the continental shelves were exposed.

Isostatic leveling causes sea level to change in relation to the local coast. If continental crust is weighted by ice during a glacial period, or if its temperature falls (density increases), it sinks. If crust loses weight (as it does when ice melts during warm periods) or warms, it rises. However, isostatic leveling is very slow.

Present-Day Oceans.

The Pacific is the largest and oldest ocean. It is ringed by subduction zones and has many volcanic islands and atolls formed at hot spots and magmatic arcs. Because few rivers drain directly into it and sediments are trapped in subduction zones and marginal seas, its seafloor has a relatively thin sediment cover.

The Atlantic Ocean is widening as lithospheric plates move apart at the Mid-Atlantic Ridge. It has few islands and broad continental shelves. Compared with the Pacific Ocean, it has more rivers and thicker average sediment cover.

The Indian Ocean is the youngest ocean. It has a complex oceanic ridge system, few islands, and thick sediment cover, especially in the north, where major rivers empty from the new, easily erodible Himalaya Mountains created at the India–Eurasia continental collision.

There are four types of marginal seas: shallow seas where continental crust is submerged, long narrow seas where continents are breaking apart, seas between continents that are moving toward a future collision, and back-arc basins behind subduction zones.

KEY TERMS

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

abyssal fans
abyssal hill
abyssal plain
asthenosphere
atoll
back-arc basin
barrier reef
basalt
buoyant
central rift valley
chaotic
coastal plain
continental collision
continental divergent plate boundary
climate
coast
continental drift
continental margins
continental shelf
continental slope
convection
convection cell
convergence
convergent plate boundary
crest
crust
deposition
dip angle
divergence
divergent plate boundary
downwell
erosion
eustasy
exotic terrane
fjords
fossils
fracture zone
fringing reef
friction
glacial
gravity
greenhouse effect
guyot
hot spot
hydrocarbon
hydrosphere
ice age
isostasy
isostatic leveling
latitudes
lithosphere
lithospheric plate
magma
magmatic arc
mantle
marginal sea
oceanic convergent plate boundary
oceanic plateau
oceanic ridge
paleomagnetic
passive margin
photosynthesis
plate tectonics
primary productivity
radioactive
rift zone
runoff
salt marshes
seafloor spreading
seamount
sediment
sedimentary arc
sedimentation
sedimentation rate
shelf break
shelf valleys
sonar
species
spreading cycle
subduct
subduction zone
submarine canyons
superplume
tomography
topographic
transform fault
transform plate boundary
trench
tsunami
turbidity
upwelled

STUDY QUESTIONS

1. Describe the Earth's mantle. How do we know what the mantle is made of and how it behaves?
2. What are the differences between continental crust and oceanic crust, and why are these differences important?
3. What is a hot-spot trail? How do hot-spot trails show that lithospheric plates move across the Earth's surface?
4. What three types of motion occur at plate boundaries?
5. List the types of convergent plate boundaries. Describe the characteristics and locations of volcanoes associated with convergent plate boundaries. Why are there few or no volcanoes at convergent plate boundaries where two continents collide?
6. What processes occur at oceanic ridges to form their mountainous topography and the fracture zones that cut across them?
7. Describe the changes in seafloor depth and sediment cover with increasing distance from an oceanic ridge. What causes these changes?
8. Why are passive margins described as "passive"? What are their characteristics?
9. Distinguish between isostatic and eustatic processes that cause sea level to change. How do these processes complicate efforts to measure changes in global sea level by measuring sea-level heights at various coastlines?

10. If there were no ice cap on the Antarctic continent, which coast of the United States—the California coast, the Pacific Northwest coast, or the Mid-Atlantic coast—would the Antarctic coast and continental shelf resemble? Why?
11. Describe and explain the principal differences in geography and seafloor topography between the Atlantic Ocean and the Pacific Ocean.

CRITICAL THINKING QUESTIONS

1. The structure of the Earth and the processes of plate tectonics depend on the way that fluids and solids of different density interact. List examples of everyday situations in which the difference in density between substances or within a substance determine their behavior. One such example is oil floating on top of water.
2. If continental crust and oceanic crust were able to flow like the material in the asthenosphere, how might the Earth be different? How would the processes of plate tectonics be different?
3. Most ice ages in the Earth's history have lasted about 50 million years. The most recent ice age began 2 to 3 million years ago. About 10,000 years ago, the glaciers that extended into much lower latitudes than they do today began to melt and the Earth's climate warmed. Does this evidence indicate that the most recent ice age is over? Explain the reasons for your answer.
4. Is it possible that plate tectonic motions occur at present on any planet or moon in our solar system other than the Earth? Why or why not?
5. Are there planets or moons on which there were probably active plate tectonic motions in the past but not at present? If so, which planets or moons are the most likely to have had such motions? Why?
6. Using the surface topography maps of the planets and moons that are now available, how would you investigate whether, in fact, there had been plate tectonic activity on these planets?
7. Volcanoes occur at intervals along the length of divergent plate boundaries, but not along the length of transform faults such as the San Andreas Fault. Why?
8. Evidence exists that there are extinct volcanoes in California arranged roughly parallel to the San Andreas Fault. What are the possible explanations for their origin? How would you determine which of the possible explanations was correct?
9. If the Atlantic Ocean stopped expanding and started to contract, what would happen to New York City? Where would it be located a few million years after this change of direction occurred? Would it still be on land? Where would the nearest mountains be? Where would the coastline be?
10. There have apparently been several spreading cycles in the Earth's history. What might cause the continents to repeatedly collect together, then break apart, only to collect together again?
11. One of the effects of the enhanced greenhouse effect is that sea level is rising and will continue to rise. What are the physical processes that will cause sea-level change as climate changes? How many of these causal factors can you list? Which of these factors would be the most important? Would any of them tend to lower sea level? What effects would a sea-level rise of 1 m or 10 m have on humans?
12. In the 1995 movie Waterworld, melting of all the ice in glaciers and the polar ice caps caused the oceans to expand and cover all the land surface on the Earth except for a few small islands. Is this possible? Why or why not?

CRITICAL CONCEPTS REMINDERS

CC1 Density and Layering in Fluids: The Earth and all other planets are arranged in layers of different materials sorted by their density.

CC2 Isostasy, Eustasy, and Sea Level: Earth's crust floats on the plastic asthenosphere. Sections of crust rise and fall isostatically as temperature changes alter their density and as their mass loading changes due to melting or to the formation of ice stemming from climate changes. This causes sea level to change on the coast of that particular section of crust. Sea level can also change eustatically when the volume of water in the oceans increases or decreases due to changes in water temperature or changes in the amount of water in glaciers and ice caps on the continents. Eustatic sea level change takes place synchronously worldwide and much more quickly than isostatic sea level changes.

CC3 Convection and Convection Cells: Fluids that are heated from below, such as Earth's mantle, or ocean water, or the atmosphere, rise because their density is reduced. They continue to rise to higher levels until they are cooled sufficiently, at which time they become dense enough to sink back down. This convection process establishes convection cells in which the heated material rises in areas of upwelling, spreads out, cools, and then sinks at areas of downwelling.

CC7 Radioactivity and Age Dating: Some elements have naturally occurring radioactive (parent) isotopes that decay at precisely known rates to become a different (daughter) isotope, which is often an isotope of another element. This decay process releases heat within the Earth's interior. Measurement of the concentration ratio of the parent and daughter isotopes in a rock or other material can be used to calculate its age, but only if none of the parent or daughter isotopes have been gained or lost from the sample over time.

CC11 Chaos: The nonlinear nature of many environmental interactions makes complex environmental systems behave in sometimes unpredictable ways. It also makes it possible for these changes to occur in rapid and unpredictable jumps from one set of conditions to a completely different set of conditions.

CC14 Phototrophy, Light, and Nutrients: Chemosynthesis and phototrophy (which includes photosynthesis) are the processes by which simple chemical compounds are made into the organic compounds of living organisms. The oxygen in Earth's atmosphere is present entirely as a result of photosynthesis.

REFERENCES

Figure 4-1 – Segar, D., and E. Stamman Segar, 1997: An eruption of Pu‘u‘ō‘ō.

Figure 4-2 – Adapted from Robertson, E. C., 2011: The Interior of the Earth. USGS, accessed on 18 June 2025, https://pubs.usgs.gov/gip/interior/.

Figure 4-3 – Adapted from E. J. Tarbuck and F. K. Lugens, 1991, *Earth Science, 6th Edition*, Macmillan Publishing Company, 528 pp.

Figure 4-7 – Adapted from Kious, W. J., and R. I. Tilling, R.I., 1996: This dynamic earth—The story of plate tectonics (online ed., ver. 1.20, 2015). U.S. Geological Survey General Interest Publication, https://doi.org/10.3133/7000097.

Figure 4-8 – Adapted from Isacks, B., J. Oliver and L. R. Sykes, 1968: Seismology and the new global tectonics. *J. Geophys. Res.*, **73**, 5855–5899, https://doi.org/10.1029/JB073i018p05855.
Adapted from Barazangi, M., and J. Dorman, 1969: World seismicity maps compiled from ESSA, Coast and Geodetic Survey, epicenter data, 1961-1967. *Bull. Seismol. Soc. Am.*, **59**, 369–380, https://doi.org/10.1785/BSSA0590010369.

Figure 4-9 – Adapted from Pipkin, B. W., 1994: *Geology and the Environment*. West Publishing Company, 466 pp.

Figure 4-9 – Adapted from Scotese, C. R. and N. M. Wright, 2018: PALEOMAP Paleodigital Elevation Models (PaleoDEMS) for the Phanerozoic. EarthByte, https://doi.org/10.5281/zenodo.5348491.

Figure 4-10 – T. H. Van Andel, 1985: *New Views of an Old Planet: Continental Drift and the History of the Earth 1 Ed*. Cambridge University Press, 324 pp.

Figure 4-11 – T. H. Van Andel, 1985: *New Views of an Old Planet: Continental Drift and the History of the Earth 1 Ed*. Cambridge University Press, 324 pp.

Figure 4-12 – Adapted from Kious, W. J., and R. I. Tilling, R.I., 1996: This dynamic earth—The story of plate tectonics (online ed., ver. 1.20, 2015). U.S. Geological Survey General Interest Publication, https://doi.org/10.3133/7000097.

Figure 4-13a – Doukas, M., 1980: July 22, 1980 eruption of Mount St. Helens sent pumice and ash 6 to 11 mi (10-18 km) into the air, and was visible in Seattle, Washington (100 mi/160 km north). USGS Volcano Hazards Program, accessed 18 June 2025, https://www.usgs.gov/media/images/july-22-1980-eruption-mount-st-helens-sent-pumice-and-ash-6.

Figure 4-13b – Harlow, D., 1991: The June 12, 1991 eruption column from Mount Pinatubo taken from the east side of Clark Air Base. USGS Volcano Hazards Program, accessed 18 June 2025, https://www.usgs.gov/media/images/june-12-1991-eruption-column-mount-pinatubo-taken-th.

Figure 4-13c – Glicken, H., 1980: Mount St. Helens on May 17, 1980, one day before the devastating eruption. The view is from Johnston's Ridge, six miles (10 kilometers) northwest of the volcano. USGS Volcano Hazards Program, accessed 18 June 2025, https://www.usgs.gov/media/images/mount-st-helens-may-17-1980-one-day-devastating-er.

Figure 4-13d – Glicken, H., 1980: MSH80 st helens from johnston ridge 09-10-80. USGS Cascades Volcano Observatory, Wikipedia, https://upload.wikimedia.org/wikipedia/commons/3/35/MSH80_st_helens_from_johnston_ridge_09-10-80.jpg, Public Domain.

Figure 4-16 – Adapted from Kious, W. J., and R. I. Tilling, R.I., 1996: This dynamic earth—The story of plate tectonics (online ed., ver. 1.20, 2015). U.S. Geological Survey General Interest Publication, https://doi.org/10.3133/7000097.

Figure 4-19 – Arctic Exploration 2002, 2002: International bathymetric chart of the Arctic Ocean, with locations of proposed expeditions of the International Survey of Arctic Ridges and Basins. NOAA/OER, accessed 24 June 2025, https://oceanexplorer.noaa.gov/explorations/02arctic/background/plan/media/arctic_bath.html.

Figure 4-20 – Adapted from Kious, W. J., and R. I. Tilling, R.I., 1996: This dynamic earth—The story of plate tectonics (online ed., ver. 1.20, 2015). U.S. Geological Survey General Interest Publication, https://doi.org/10.3133/7000097.

Figure 4-24 – Segar, D., and E. Stamman Segar, 1997: The 1983 eruption of Pu‘u‘ō‘ō.

Figure 4-25a – Clague, D. A., and G. B. Dalrymple, 1987: The Hawaiian-Emperor volcanic chain. Volcanism in Hawaii, Part 1, R. W. Decker, T. L. Wright, and P. H. Stauffer, Eds. 5–100, https://pubs.usgs.gov/pp/1987/1350/pdf/chapters/pp1350_ch1.pdf.

Figure 4-27 – Adapted from Dewey, J. F. and J. M. Bird, 1970: Plate tectonics and geosynclines. Tectonophysics, 10, 625–638. https://doi.org/10.1016/0040-1951(70)90050-8.

Figure 4-28 – NASA World Wind, 2006: Orona Atoll. https://commons.wikimedia.org/wiki/File:Orona_Atoll.png#/media/File:Orona_Atoll.png.

Figure 4-29 – Adapted from P.R. Pinet, 1992: *Oceanography, an introduction to the planet Oceanus*. West Publishing Company, 350 pp.

Figure 4-30 – NASA, Johnson Space Center, 2003: Bora Bora ISS006. https://en.m.wikipedia.org/wiki/File:Bora_Bora_ISS006.jpg.

Figure 4-31 – Adapted from Van Andel, T. H., 1985: *New views on an old planet, continental drift and the history of earth*. Cambridge University, 324 pp.

Figure 4-32a – Data from Cross-Chapter Box 2.1, Figure 1 in IPCC, 2021: Chapter 2. In: Climate Change 2021: The Physical Science Basis. V. Masson-Delmotte, and Coauthors, Eds. Cambridge University Press, 287–422, https://doi.org/10.1017/9781009157896.004.

Figure 4-32b – Data from Bintanja, R., R.S.W. van de Wal, and J. Oerlemans. 2005. Modelled atmospheric temperatures and global sea levels over the past million years. *Nature*, **437**, 125–128, https://doi.org/10.1038/nature03975.

Figure 4-32c – Data from Bintanja, R., R.S.W. van de Wal, and J. Oerlemans. 2005. Modelled atmospheric temperatures and global sea levels over the past million years. *Nature*, **437**, 125–128, https://doi.org/10.1038/nature03975.

Figure 4-34a – Data from Bintanja, R., R.S.W. van de Wal, and J. Oerlemans. 2005. Modelled atmospheric temperatures and global sea levels over the past million years. *Nature*, **437**, 125–128, https://doi.org/10.1038/nature03975.

Figure 4-34b – Data from Bintanja, R., R.S.W. van de Wal, and J. Oerlemans. 2005. Modelled atmospheric temperatures and global sea levels over the past million years. *Nature*, **437**, 125–128, https://doi.org/10.1038/nature03975.

Figure 4-34c – Data from GISTEMP Team, 2025: GISS Surface Temperature Analysis (GISTEMP), version 4. NASA Goddard Institute for Space Studies. Dataset accessed 24 June 2025, https://data.giss.nasa.gov/gistemp/.

Figure 4-34c – Data from Willis, J. K., and Coauthors, 2025: NASA-SSH Global Mean Sea Level from Simple Gridded Sea Surface Height. Ver. 1. Dataset accessed 24 June 2025, https://doi.org/10.5067/NSIND-GMSV1.