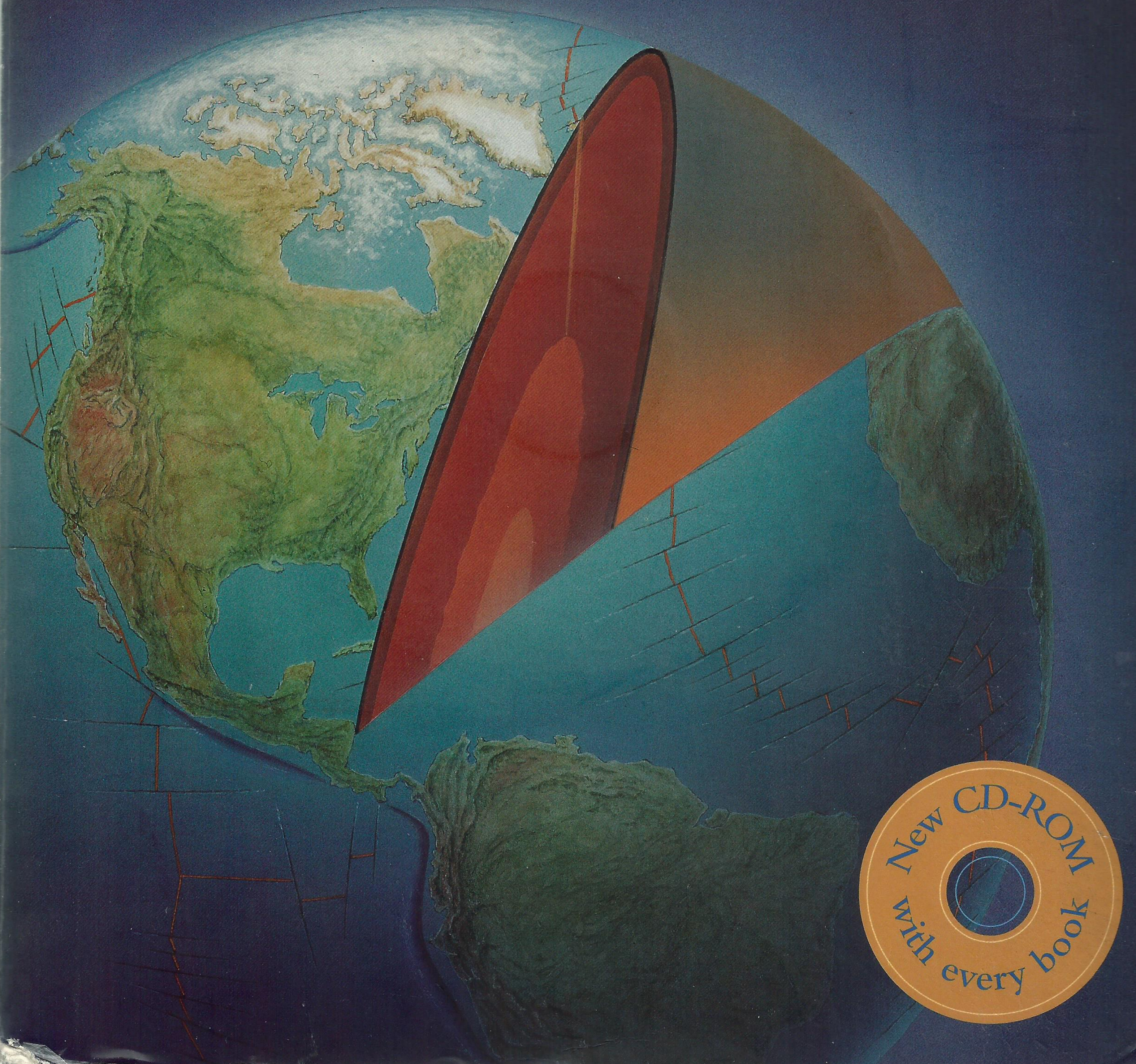
Understanding Earth

SECOND EDITION

Frank Press Raymond Siever



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RAYMOND SIEVER
Harvard University

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TO OUR CHILDREN, AND OUR CHILDREN'S CHILDREN;
MAY THEY LIVE IN HARMONY WITH EARTH'S ENVIRONMENT.

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The Oceans

For most of human history, the 71 percent of Earth's surface covered by the oceans was a mystery. The large populations who lived at the edge of the sea knew well the force of the waves and the rise and fall of tides, but they could only guess at the nature of the seafloor deeper than the shallowest coastal waters. By the middle of the nineteenth century, occasional oceangoing ships made scientific observations of water depths, the plants and animals of the sea, and the chemistry of seawater. Then, in 1872, H.M.S. *Challenger,* a small wooden British warship converted and fitted out for the first specifically scientific study of the seas, left England for a four-year voyage over the world's oceans. The 50 thick volumes of reports from that expedition gave the public its first knowledge of the Mid-Atlantic Ridge, one of the longest, highest mountain ranges in the world—all of it underwater. The *Challenger* expedition discovered great areas of submerged hills and flat plains, extraordinarily deep trenches, and submarine volcanoes.

Today, more than a century after that pioneering voyage, hundreds of oceanographic research vessels from many countries ply the seas in search of answers to the questions first raised by the early discoveries. What tectonic forces raised the submarine mountain ranges and depressed the trenches? Why are some areas flat and others hilly? Although oceanographers made many important discoveries in the first half of this century, the answers to most of these questions had to await the plate-tectonics revolution of the late 1960s. In fact, it was geological and geophysical observations of the ocean floors, not the continents, that led to the theory of plate tectonics.

In this chapter, we examine what these early and more modern researchers have discovered about Earth's oceans, with their waves and tides; submerged mountains and valleys; underwater volcanoes; and many kinds of rocks, sediments, and chemical components. First, though, we define some terms.

We refer to the oceans both as the five major oceans (Atlantic, Pacific, Indian, Arctic, and Antarctic) and as the single connected body of water called the **world ocean.** The term *sea* includes both the oceans and smaller bodies of water set off somewhat from the world ocean. Thus, the Mediterranean Sea is narrowly connected with the Atlantic Ocean by the Straits of Gibraltar and with the Indian Ocean by the Suez Canal. Other seas, such as the North Sea and the Atlantic Ocean, are broadly connected. In the world ocean, seawater—the salty water of the oceans and seas—is remarkably constant in its general chemical composition from year to year and from place to place. This is so because the oceans maintain an equilibrium determined by the general composition of river waters entering the sea, the composition of the sediment brought into the oceans, and the formation of new sediment in the ocean (see Chapter 24).

We begin our exploration of the oceans with shorelines, where we can observe the constant motion of ocean waters and their effects on the shore.

The Edge of the Sea: Waves and Tides

Coasts, the broad regions where land meets sea, present striking contrasts of landscape. On the coast of North Carolina, for example, long, straight, sandy beaches stretch for miles along low coastal plains (Figure 17.1). In New England, by contrast, rocky cliffs bound elevated shores, and the few beaches that occur are made of gravel (Figure 17.2). Many of the seaward edges of islands in the tropics, such as those in the Caribbean Sea, are coral reefs, the delight of divers. As we will see, tectonics, erosion, and sedimentation work together to create this great variety of shapes and materials.

FIGURE **17.1** A long, straight, sandy beach on South Pea Island, North Carolina. *(Peter Kresan.)*

FIGURE 17.2 Small pocket beach *(right foreground)* of pebbles and cobbles. Acadia National Park, Maine. *(Ric Ergenbright Photography.)*

The major geological forces operating at the **shoreline,** the line where the water surface intersects the shore, are waves and tides. Together, they erode even the most resistant rocky shores. Waves and tides create currents, which transport sediment produced by erosion of the land and deposit it on beaches and in shallow waters along the shore.

Wave Motion: The Key to Shoreline Dynamics

Centuries of observation have taught us that waves are changeable. During quiet weather, waves roll regularly into shore with calm troughs between them. In the high winds of a storm, however, waves are everywhere, moving in a confusion of shapes and sizes; they may be low and gentle far from the shore, yet become high and steep as they approach land. High waves can break on the shore with fearful violence, shattering concrete seawalls and tearing apart houses built along the beach. To understand the dynamics of shorelines and to make sensible decisions about shore development, we need to understand how waves work. The behavior of waves has long been observed by seafarers on sailing ships, who wanted to know which wave conditions would speed their way and which could endanger the ship.

Waves are created by the wind blowing over the surface of the water, transferring the energy of motion from air to water. As a gentle breeze of 5 to 20 km per hour starts to blow over a calm sea surface, ripples—little waves less than a centimeter high—take shape (see Table 14.1). As the speed of the wind increases to about 30 km per hour, the ripples grow to full-sized waves. Stronger winds create larger waves and blow off their tops to make whitecaps. The height of the waves increases as

- The wind speed increases.
- The wind blows for longer times.
- The distance over which the wind blows the water increases.

Storms blow up large, irregular waves that radiate outward from the storm area, like the ripples moving outward from a pebble dropped into a still pond. As the waves travel out from the storm center in ever-widening circles, they become more regular, changing to low, broad, rounded waves called **swell,** which can travel hundreds of kilometers. Several storms at different distances from a shoreline, each producing its own pattern of swell, account for the often irregular intervals between waves approaching the shore.

Waves travel as a form, but the water stays in the same place. If you have seen waves in an ocean or a large lake, you have probably noticed how a piece of wood or other light material floating on the water moves a little forward as the top of a wave passes and then a little backward as the trough between waves passes. While moving back and forth, the wood stays in roughly the same place, and so does the water around it.

Small water particles at the surface or beneath the waves move in circular vertical orbits. At any given point along the path of a wave, all the water particles are at the same relative positions in their orbits regardless of their depth. The radii of the orbits are large near the water surface, but they gradually decrease to

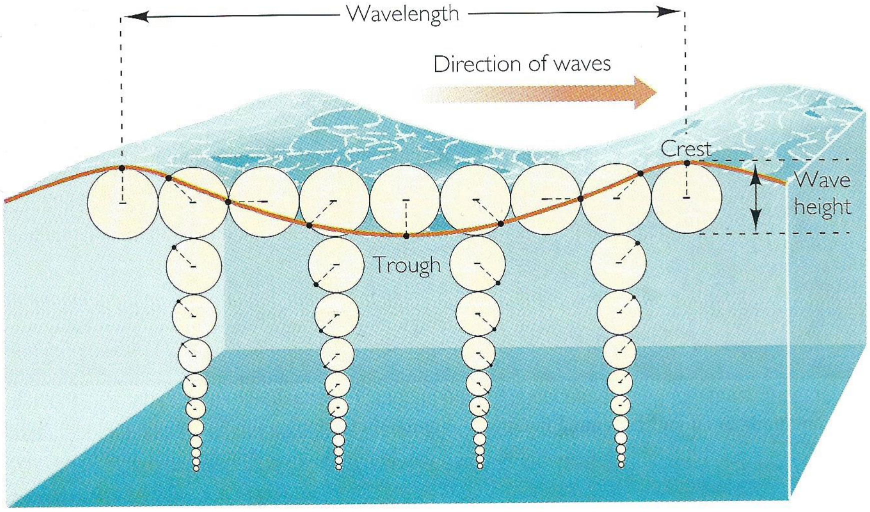


FIGURE 17.3 The orbital movement of water particles as a wave advances. Water particles at the surface or beneath waves move in circular vertical orbits. Note that orbits decrease in radius with depth. The forward movement of the wave form (solid line) is traced out by the paths of the orbits, the troughs following the bottoms of the orbits and the crests following the tops of the orbits.

zero at some depth below, as shown in Figure 17.3. This is so because the greater the distance the water particle is from the surface, the farther it is from the force of the wind. And the farther the particle is from the force of the wind, the more important the viscous forces resisting water movement become. A wave form is made as many water particles move to the top of the orbit, and the wave advances as the particles continue around the orbit. The trough is created as the particles reach the bottom of the orbit.

We describe a wave form in terms of three characteristics (see Figure 17.3):

- **Wavelength,** the distance between crests
- **Wave height,** the vertical distance between the crest and the trough
- **Period,** the time it takes for successive waves to pass

We measure the velocity at which a wave moves forward by using a basic equation:

$$V = \frac{L}{T}$$

where V is the velocity, L is the wavelength, and T is the period. Thus, a typical wave with a length of 24 m and a period of 8 seconds would have a velocity of 3 m per second. The periods of most waves range from just a few seconds to as long as 15 or 20 seconds, with wavelengths varying from about 6 m to as much as 600 m. Consequently, wave velocities vary from 3 to 30 m per second. At a depth of about one-half the wavelength, orbital motion stops and wave motion ceases. That is why deep divers and submarines are unaffected by the waves at the surface.

The Surf Zone

Swell becomes higher as it approaches the shoreline. There it assumes the familiar sharp-crested wave shape. These waves are called *breakers* because, as the waves come closer to shore, they break and form **surf,** a foamy, bubbly surface. The offshore belt, along which breaking waves collapse as they approach the shore, is the **surf zone.** Breaking waves pound the shore, eroding and carrying away sand, weathering and breaking up solid rock, and destroying structures built close to the shoreline.

The transformation from swell to breakers starts where the bottom shallows to less than one-half the wavelength of the swell. At that point, the small orbital motions of the water particles just above the bottom become restricted because the water can no longer move vertically. Right next to the bottom, the water can only move back and forth horizontally. Above that, the water can move vertically just a little, combining with the horizontal motion to give a flat elliptical orbit rather than a circular one (Figure 17.4). The orbits become more circular the farther they are from the bottom.

The change from circular to elliptical orbits slows the whole wave, because the water particles take longer to travel around ellipses than around circles. While the wave slows, its period remains the same because the swell keeps coming in from deeper water at the same rate. From the wave equation, we

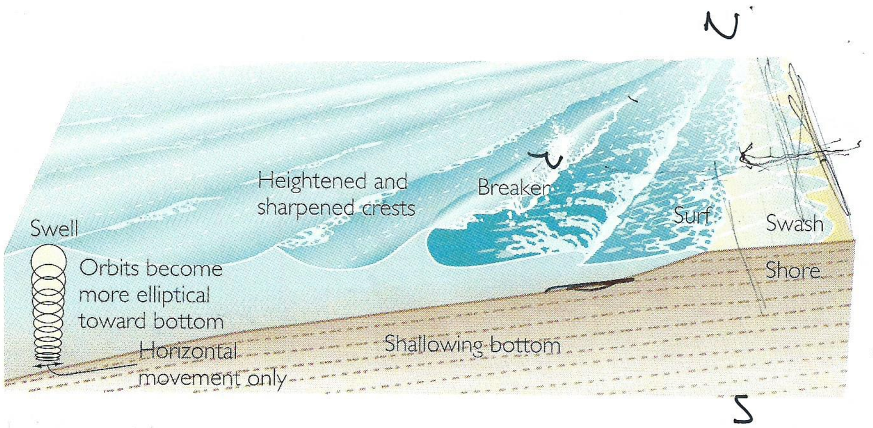


FIGURE 17.4 Formation of a breaking wave as swell meets a shallowing bottom. Note that the orbits of water particles become more elliptical as they approach a shallow bottom. As water particles take longer to travel around their ellipses, the whole wave slows. Waves become more closely spaced until, heightened and sharpened, they break with a crash in the surf zone.

know that if the velocity decreases and the period remains constant, the wavelength must also decrease. The typical wave we used as our example earlier might, while keeping the same period of 8 seconds, change to a length of 16 m and thus a velocity of 2 m per second. Thus, the waves become more closely spaced, higher, and steeper, and their wave crests become sharper.

As a wave rolls toward the shore, it becomes so steep that the water can no longer support itself, and the wave breaks with a crash in the surf zone (see Figure 17.4). Gently sloping bottoms cause the waves to break farther out, and steeply sloping bottoms make waves break closer to shore. Where rocky shores are bordered by deep water, the waves break directly on the rocks with a force amounting to hundreds of tons per square meter, throwing water high into the air. It is not surprising that concrete seawalls built to protect buildings along the shore quickly start to crack and must be repaired constantly.

After breaking at the surf zone, the waves, now reduced in height, continue to move in, breaking again right at the shoreline (Figure 17.5). They run up onto the sloping front of the beach, forming an uprush of water called **swash.** The water then runs back down again as **backwash.** Swash can carry sand and, if the waves are high enough, large pebbles and cobbles. The backwash carries the particles back down again.

The motion of the water back and forth near the shore is strong enough to carry sand grains and even gravel. Fine sand can be moved by wave action in water up to about 20 m deep. Large waves caused by intense storms can scour the bottom at much greater depths, down to 50 m or more.

FIGURE 17.5 Waves obliquely approaching the shoreline and being refracted to be almost parallel to the shoreline. The white color shows the surf zone. Oceanside, California. *(John S. Shelton.)*

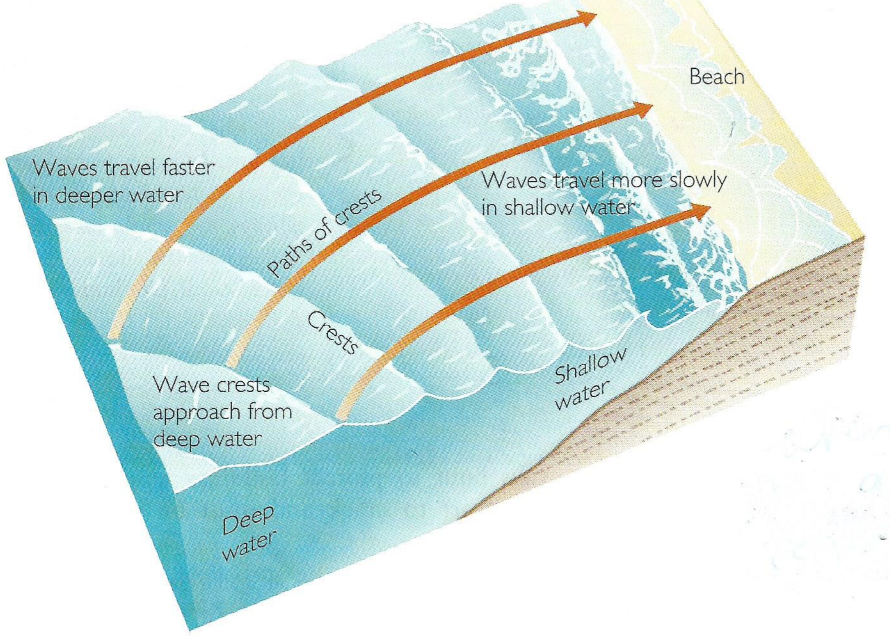


FIGURE 17.6 Wave refraction is the bending of lines of wave crests as they approach the shore from an angle.

Wave Refraction

Far from shore, the lines of swell are parallel to one another but are usually at some angle to the shoreline. As the waves approach the beach over a shallowing bottom, the rows of waves gradually bend to a direction more parallel to the shore (Figure 17.6). This bending of lines of wave crests as they approach the shore from an angle is called **wave refraction.** It is similar to the bending of light rays in optical refraction, which makes a pencil half in and half out of water appear to bend at the water surface. Water refraction begins as a wave approaches the shore at an angle. The part of the wave closest to the shore encounters the shallowing bottom first, and the orbits of the water particles in that part of the wave become more elliptical. As this happens, the front of the wave slows. Then the next part of the wave meets the bottom and also slows. Meanwhile, the parts closest to shore have moved into even shallower water and slowed even more. Thus, in a continuous transition along the wave crest, the line of waves bends toward the shore as it slows.

Wave refraction results in more intense wave action on projecting headlands and less intense action in indented bays, as Figure 17.7 illustrates. The water becomes shallow more quickly around headlands than in the surrounding deeper water on either side. Waves are refracted around headlands—that is, they are bent toward the projecting part of the shore from both sides. The waves converge around the point of land and expend proportionately more of their energy breaking there than at other places along the shore. Thus, erosion by waves is concentrated at headlands and tends to wear them away more quickly than it does straight sections of shoreline.

The opposite happens as a result of wave refraction in a bay. The waters in the center of the bay are deeper, so the waves are refracted on either side into shallower water. The energy of wave motion is diminished at the center of the bay, making bays good harbors for ships.

Although refraction makes waves more parallel to the shore, many waves still approach at some small angle. As the waves break on the shore, the swash moves up the beach slope perpendicular to this small angle. The backwash runs down the slope at a similar small angle but in the opposite direction, like a parabola. The combination of the two motions results in a trajectory that moves the water a short way down the beach (Figure 17.8). Sand grains carried by swash and backwash are thus moved along the beach in a zigzag motion known as **longshore drift.**

Waves approaching the shoreline at an angle can also cause a **longshore current,** a shallow-water current that is parallel to the shore. The water that moves with swash and backwash in and out from the shore at an angle creates a zigzag path of water particles that adds up to a net transport along the shore

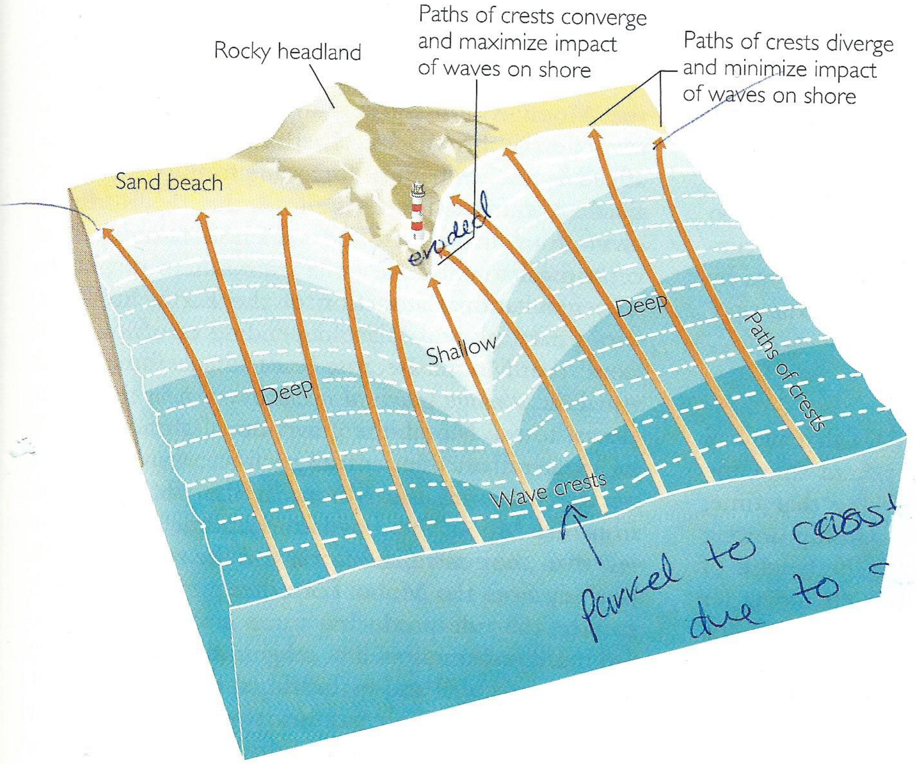


FIGURE 17.7 Wave refraction around a headland and bay. Wave energies are concentrated at headlands and dispersed at bays.

in the same direction as the longshore drift. Much of the net transport of sand along many beaches comes from this kind of current. Longshore currents are prime determiners of the shapes and extent of sand bars and other depositional shoreline features. At the same time, because of their ability to erode loose sand, longshore currents may remove much sand from a beach. Longshore drift and longshore currents working together are potent processes in the transport of large amounts of sand on beaches and in very shallow waters. In deeper but still shallow waters (less than 50 m), longshore currents—especially those running during large storms—strongly affect the bottom.

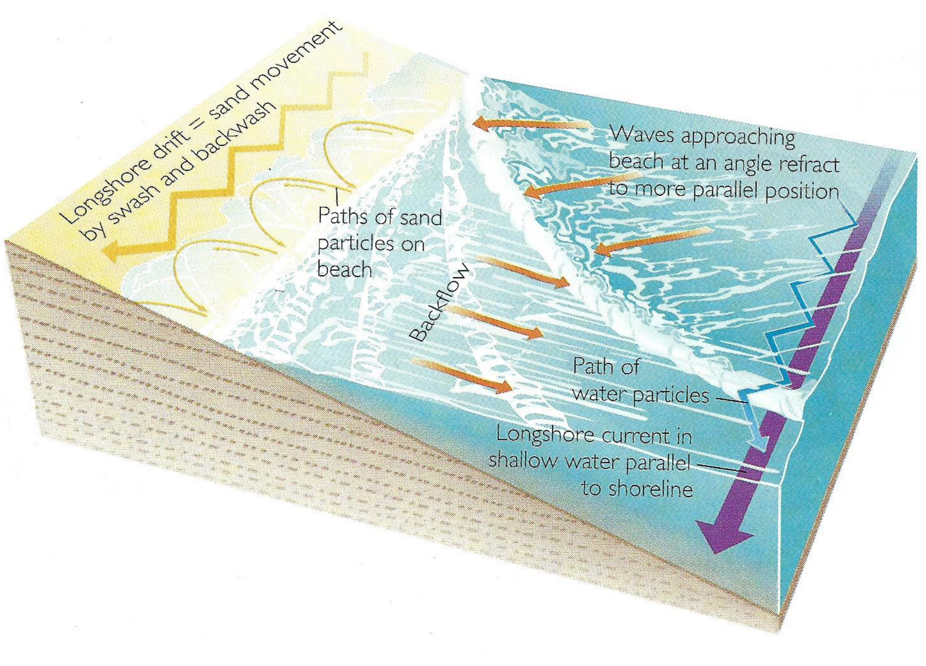


FIGURE 17.8 Longshore drift is the zigzag movement of sand grains thrown up on the shore by waves approaching at an angle. Waves approaching the shore obliquely also cause a zigzag movement of water particles in shallow water, giving rise to a longshore current.

Some types of currents related to longshore currents can pose a threat to unwary swimmers. A *rip current,* for example, is a strong flow of water moving perpendicularly outward from the shore. It occurs when a longshore current builds up along the shore and the water piles up imperceptibly until a critical point is reached. At that point, the water breaks out to sea, flowing through oncoming waves in a fast current. Swimmers can avoid being carried out to sea by swimming parallel to the shore to get out of the rip.

The Tides

The twice-daily rise and fall of the sea that we call **tides** have been known to mariners and shoreline dwellers for thousands of years. During that time, many observers also noticed a relationship among the position and phases of the Moon, the heights of the tides, and the times of day at which the water reaches high tide. Not until the seventeenth century, however, when Isaac Newton formulated the laws of gravitation, did we begin to understand that the tides result from the gravitational pull of the Moon and the Sun on the water of the oceans.

THE MOON, THE SUN, GRAVITY, AND THE TIDES The Earth and the Moon attract each other strongly with a gravitational force that is slightly greater on the sides of the bodies that face each other. The gravitational attraction between any two bodies decreases as they get farther apart. Thus, the tide-producing force varies on different parts of the Earth, depending on whether they are closer to or farther from the Moon.

Because the average distance between Earth and the Moon is constant over time, the gravitational attraction of the Moon to Earth must be exactly balanced by the centrifugal force due to the rotation of Earth. (Centrifugal force is a force that repels a body away from an axis around which it rotates.) This centrifugal force is the same everywhere on Earth. The gravitational attraction of the Earth, however, varies from place to place depending on its distance from the Moon. Because centrifugal force must equal the average gravitational attraction, at the point closest to the Moon, the gravitational attraction will be greater than the centrifugal force, resulting in a net force toward the Moon. At the point farthest from the Moon, the gravitational attraction will be less than the centrifugal force, and the net force will be away from the Moon. For other points on Earth's surface, the tide-producing force is the vector sum (taking direction as well as magnitude into account) of the centrifugal and gravitational forces.

The net tidal force causes two bulges of water on the Earth's oceans, one from the side nearest the Moon, where the net force is toward the Moon, and the other on the side farthest from the Moon, where the net force is away from the Moon (Figure 17.9). The net gravitational attraction between the oceans and the Moon is at a maximum on the side of Earth facing the Moon and at a minimum on the side facing away from the Moon. As Earth rotates, the bulges of water stay approximately aligned. One

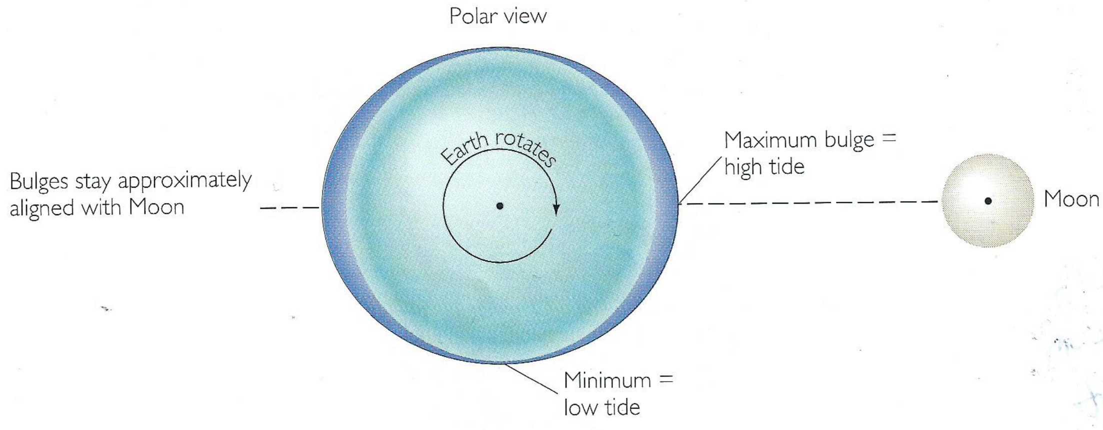


FIGURE 17.9 The Moon's gravitational attraction causes two bulges of water on the Earth's oceans, one on the side nearest the Moon and the other on the side farthest from the Moon. As the Earth rotates, these bulges remain aligned and pass over Earth's surface, forming the high tides.

always faces the Moon, the other is always directly opposite. These bulges passing over the rotating Earth are the high tides.

The Sun, although much farther away, has so much mass (and thus so much gravity) that it, too, causes tides. Sun tides are a little less than half the height of Moon tides. Sun tides are not synchronous with Moon tides (Figure 17.10). Sun tides come as the Earth rotates once every 24 hours, the length of a solar day. The rotation of the Earth with respect to the Moon is a little longer because the Moon is moving around the Earth, giving a lunar day of 24 hours and 50 minutes. In that lunar day, there are two high tides, with two low tides between them.

When the Moon, Earth, and Sun line up (see Figure 17.10a), the gravitational pulls of the Sun and the Moon reinforce each other. This produces the **spring tides,** which are the highest tides; they derive their name from their height, not from the season. They appear every two weeks at full and new Moon. The lowest tides, the **neap tides,** come in between, at first- and third-quarter Moon, when the Sun and Moon are at right angles to each other with respect to the Earth (see Figure 17.10b).

Although the tides occur regularly everywhere, the difference between high and low tides varies in different parts of the ocean. As the tidal bulges of water rise and fall, they also move along the surface of the ocean, encountering obstacles, such as continents and islands, that hinder the flow of water. In the middle of the Pacific Ocean—in Hawaii, for example, where there is minimal constriction and obstruction of the flow of the tides—the difference between low and high tides is only 0.5 m. On the Pacific coast near Seattle, where the coast along Puget Sound is very irregular, the tides are constricted through narrow passageways, and the difference between the two tides is about 3 m. Extraordinary tides occur in a few places, such as the Bay of Fundy in eastern Canada, where the tidal range can be more than 12 m. Because many people living along the shore need to know when tides will occur, governments publish tide tables showing predicted tide heights and times; these tables combine local experience with knowledge of the astronomical motions of Earth and the Moon with respect to the Sun.

Tides may combine with waves to cause extensive erosion of the shore and destruction of shoreline property. Intense storms passing near the shore during a spring tide may produce **tidal surges,** waves at high tide that can overrun the entire beach and batter sea cliffs. Tidal surges are not to be confused with the commonly but incorrectly termed "tidal waves." Although there are no such waves associated with the tides, there are unusually large ocean waves called *tsunamis* (a Japanese word), which are caused by undersea events such as earthquakes, landslides, and the explosion of oceanic volcanoes (see Figure 18.20 and Feature 18.1).

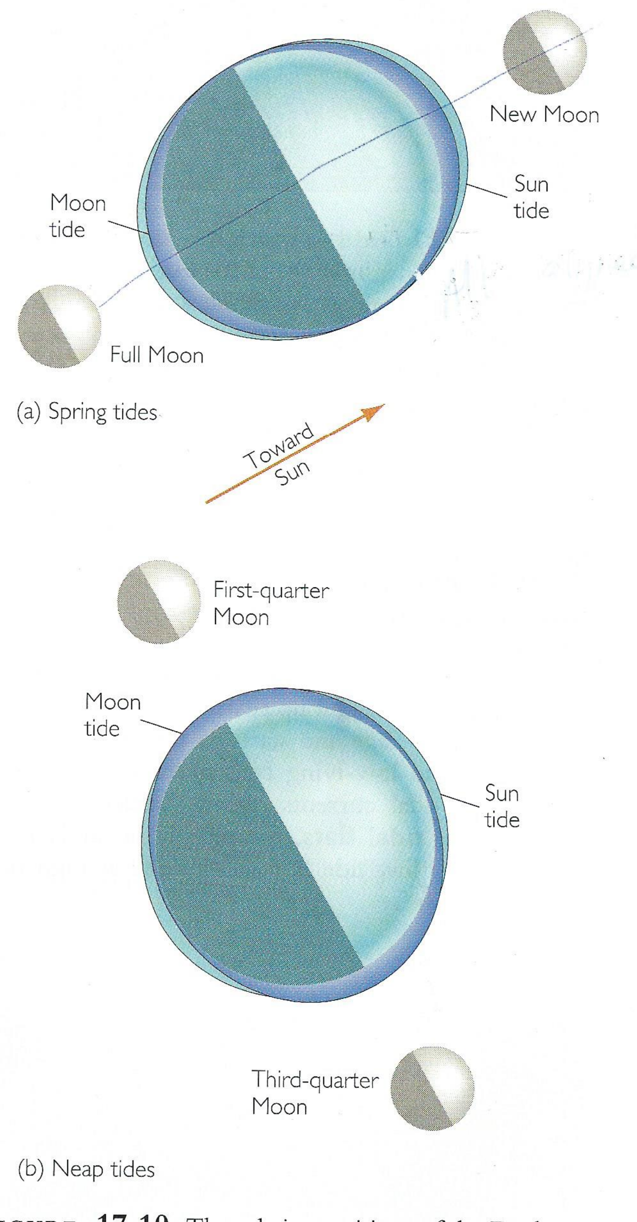


FIGURE 17.10 The relative positions of the Earth, Moon, and Sun determine the heights of high tide during the lunar month. (a) At new and full Moon, Sun and Moon tides reinforce each other and make the highest (spring) high tides. (b) At first- and third-quarter Moon, Sun and Moon tides are in opposition, causing low (neap) tides.

FIGURE 17.11 Tidal flats, such as this one at Mont-Saint-Michel, France, may be extensive areas covering many square kilometers but most often are narrow strips seaward of the beach. When a very high tide advances on a broad tidal flat like Mont-Saint-Michel's, it may move so rapidly that areas are flooded faster than a person can run. The beachcomber is well advised to learn the local tides before wandering. *(Thierry Prat/Sygma.)*

Tidal Currents Tides moving near shorelines cause currents that can reach speeds of a few kilometers per hour. As the tide rises, the water flows in toward the shore as a **flood tide,** moving into shallow coastal marshes and up small streams. As the tide passes the high stage and starts to fall, the **ebb tide** moves out, and low-lying coastal areas are exposed again. Such tidal currents meander across and cut channels into **tidal flats,** the muddy or sandy areas that lie above low tide but are flooded at high tide (Figure 17.11).

Shorelines

Waves, longshore currents, and tidal currents interact with the rocks and tectonics of the coast to shape shorelines into a multitude of forms. We can see these factors at work in the most popular of shorelines, beaches.

Beaches

A beach is a shoreline made up of sand and pebbles. Beaches may change shape from day to day, week to week, season to season, and year to year. Waves and tides sometimes broaden and extend a beach by depositing sand and sometimes narrow it by carrying sand away.

Many beaches are straight stretches of sand that range from a kilometer to more than a hundred kilometers long; others are smaller crescents of sand between rocky headlands. Belts of dunes border the landward edge of many beaches; bluffs or cliffs of sediment or rock border others. Beaches may have tide terraces—flat, shallow areas between the upper beach and an outer bar of sand—on their seaward sides (Figure 17.12).

The Structure of Beaches

Figure 17.13 shows the major parts of a beach, all of which may not be present at all times on any particular beach. Farthest out is the **offshore,** bounded by the surf zone, where the bottom begins to become shallow enough for waves to break. The **foreshore** includes the surf zone; the tidal flat; and, right at the shore, the swash zone, a slope dominated by the

FIGURE 17.12 Tide terrace. At low tide, the outer ridge (a sandbar at high tide) is exposed. Also exposed is the shallow depression between the ridge and the upper beach, which is rippled by the tidal flow in many places. *(James Valentine.)*

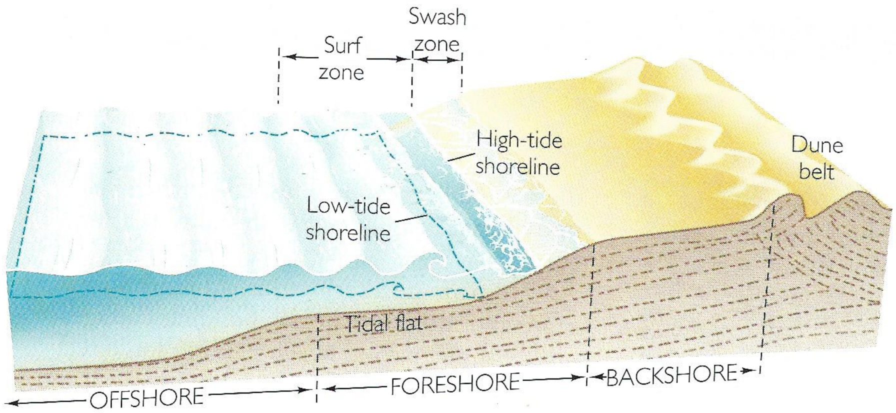


FIGURE 17.13 A profile of a beach, showing its major parts.

swash and backwash of the waves. The **backshore** extends from the swash zone up to the highest level of the beach.

The Sand Budget of a Beach A beach is a scene of incessant movement. Each wave moves sand back and forth with swash and backwash. Both longshore drift and longshore currents move sand down the beach. At the end of a beach and to some extent along it, sand is removed and deposited in deep water. In the backshore or along sea cliffs, sand and pebbles are freed by erosion and replenish the beach. The wind that blows over the beach transports sand, sometimes offshore into the water and sometimes onshore onto the land.

All these processes together maintain a balance between adding and removing sand, resulting in a beach that may appear to be stable but is actually exchanging its material on all sides. The sand budget of a beach—the inputs and outputs by erosion and sedimentation—is illustrated in Figure 17.14. At any point along the stretch, the beach gains sand by the inputs: from erosion of material along the backshore; from longshore drift and longshore current; and from rivers that enter the sea along the shore, bringing in sediment. The beach loses sand from the out-

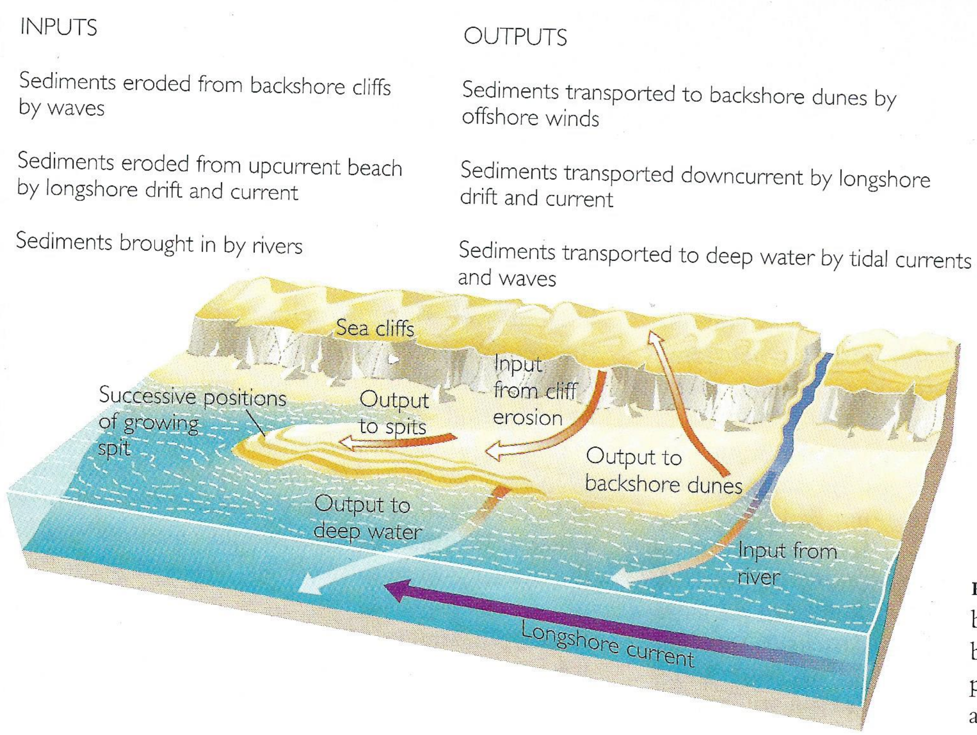


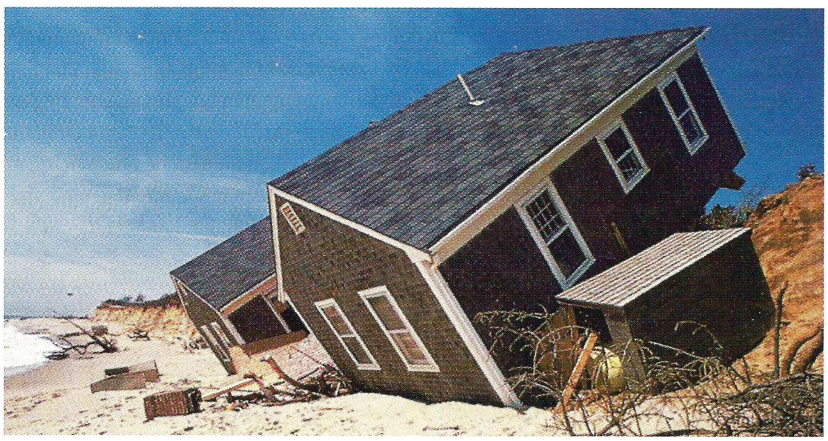
FIGURE 17.14 The beach budget is a balance between inputs and outputs of sand by erosion and sedimentation.

17.1 LIVING ON EARTH

Preserving Our Beaches

Orrin Pilkey of Duke University, a geologist and oceanographer, is in the forefront of scientists concerned about saving our beaches and halting massive development on fragile shorelines. Battered by the waves eroding the shoreline, many houses could be saved by building concrete buttresses, seawalls, and other structures designed to save shoreline property, but these structures would destroy the beach. Pilkey, a well-known researcher on coastal processes, is an advocate for the beaches of the Carolinas, which have come under heavy pressure from commercial developers. Knowing how the beach system works, he believes it is foolish to try to interfere with the natural process by which beaches remain in dynamic equilibrium with the waves and currents.

Humans are altering this equilibrium on more and more beaches by building cottages on the shore; paving beach parking lots; erecting seawalls; and constructing groins, piers, and breakwaters. A *groin* is a barrier built out from shore to prevent sand movement and beach erosion. The consequence of poorly thought-out construction is the shrinkage of the beach in one place and its growth in another, usually where no one wants it. The classic example is a narrow groin built out from shore at right angles to it. Over the following months and



A house tilts as sand is eroded from the beach and cliffs *(left background)* by storm waves. Nantucket, Massachusetts. *(Steve Rose/Rainbow.)*

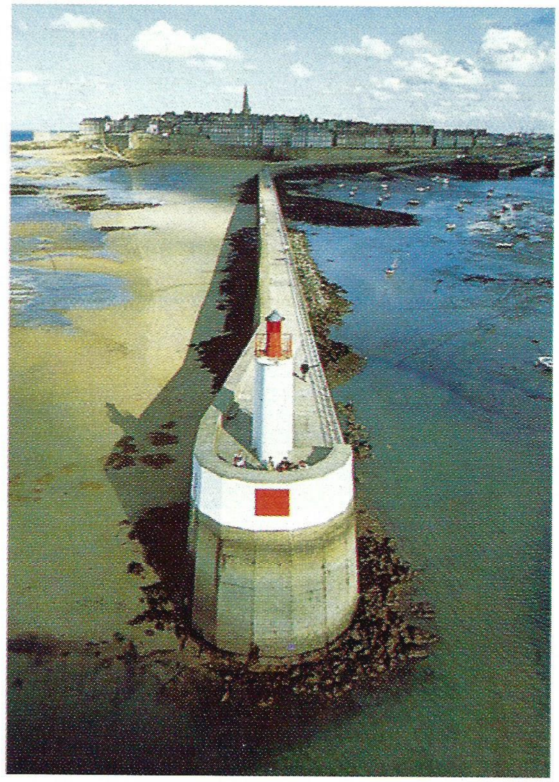
puts: winds carry sediments to backshore dunes, longshore drift and current carry it downcurrent, and deep water transports it by currents and waves during storms.

If the total input balances the total output, the beach is in equilibrium and keeps the same general form. If input and output are not balanced, the beach either grows or shrinks. Temporary imbalances are natural over weeks, months, or years. A series of large storms, for example, might move large amounts of sand from the beach to somewhat deeper waters on the far side of the surf zone, narrowing the beach. Then, in a slow return to equilibrium over weeks of mild weather and low waves, the sand might move onto shore and rebuild a wide beach. Without this constant shifting of the sands, beaches might be unable to recover from trash, litter, and some kinds of pollution. Within a year or two, even oil from spills will be transported or buried out of sight, although the tarry residue may later be uncovered in spots. Beaches would clean up rapidly if the littering were to stop.

Some Common Forms of Beaches We can now account for some common beaches. Long, wide, sandy beaches grow where sand inputs are years, the sand disappears from the beach on one side and greatly enlarges the beach on the other side—much to the surprise of the builders. As landowners and developers bring suit against one another and against state governments, trial lawyers take the issue of "sand rights"—the beach's right to the sand it naturally contains—into the courts.

The disappearance and enlargement of beaches are the predictable results of a longshore current. The waves, current, and drift bring sand toward the groin from the upcurrent direction (usually the dominant wind direction). Stopped at the groin, they dump the sand there. On the downcurrent side of the groin, the current and drift pick up again and erode the beach. On this side, however, replenishment of sand is sparse because the groin blocks the current. As a result, the beach budget is out of balance, and the beach shrinks. If the groin is removed, the beach relaxes to its former state.

The only way to save a beach is to leave it alone. Even if concrete walls and piers can be kept in repair with large expenditures of money, many times at public expense, the beach itself will suffer. Along some beaches, resort hotels truck in sand to replace that lost, but that expensive solution is temporary, too. Sooner or later, we must learn to let the beaches remain in their natural state.



Construction of groins along a shore to control erosion of a beach may produce erosion downcurrent of the groin and loss of parts of the beach *(right of groin)* while sand piles up on the other side *(left of groin)*. Longshore current flows from left to right. *(Philip Plisson/Explorer.)*

abundant, often where soft sediments make up the coast. Where the backshore is low and the winds blow from onshore, wide dune belts border the beach. If the shoreline is tectonically elevated and the rocks are hard, cliffs line the shore, and any small beaches that evolve are composed of material eroded from the cliffs. Where the shore is low-lying, sand is abundant, and tidal currents are strong, extensive tidal flats are laid down and are exposed at low tide.

What happens if one of the inputs is cut off—for example, by a concrete wall built at the top of the beach to prevent erosion? Because erosion supplies sand to the beach as one of the inputs, preventing it cuts the sand supply and so shrinks the beach. Attempts to save the beach may actually destroy it (see Feature 17.1).

Erosion and Deposition at Shorelines

The topography of the shoreline, like that of the interior, is a product of tectonic forces elevating or depressing the Earth's crust, erosion wearing it down, and sedimentation filling in the low spots. Thus, the factors directly at work are

FIGURE 17.15 The Twelve Apostles, a group of stacks at Port Campbell, Australia, developed in horizontal beds of sedimentary rock. These remnants of shore erosion are left as the shoreline retreats. *(Kevin Schafer.)*

- Uplift of the coastal region, which leads to erosional coastal forms
- Subsidence of the coastal region, which produces depositional coastal forms
- The nature of the rocks or sediments at the shoreline
- Changes in sea level, which affect the drowning or emergence of a shoreline
- The average and storm wave heights
- The heights of the tides, which affect both erosion and sedimentation

Erosional Coastal Forms Erosion is active at tectonically uplifted rocky coasts. Along these coasts, prominent cliffs and headlands jut into the sea, alternating with narrow inlets and irregular bays with small beaches. Waves crash against rocky shorelines, undercutting cliffs and causing huge blocks to fall into the water, where they are gradually worn away. As the sea cliffs retreat by erosion, isolated remnants called **stacks** are left standing in the sea, far from the shore (Figure 17.15). Erosion by waves also planes the rocky surface beneath the surf zone and creates a **wave-cut terrace,** sometimes visible at low tide (Figure 17.16). Wave erosion continuing over long periods may straighten shorelines, as headlands retreat faster than recesses and bays.

Where relatively soft sediments or sedimentary rocks make up the coastal region, the slopes are gentler and the heights of shoreline bluffs are lower. Waves efficiently erode these softer materials; erosion of bluffs on such shores may be extraordinarily rapid. The high sea cliffs of soft glacial materials along the Cape Cod National Seashore in Massachusetts, for instance, are retreating about a meter each year. Since Henry David Thoreau walked the

FIGURE 17.16 Wave-cut terrace at Bolinas Point, California. The cliffs have retreated from left to right as erosion planed a terrace at low-tide level. *(John S. Shelton.)*

FIGURE 17.17 Aerial view of the southern tip of Cape Cod, Monomoy Point, Massachusetts. This spit has advanced into deep water to the south *(foreground)* from the main body of the Cape to the north *(background)*. *(Steve Dunwell/The Image Bank.)*

entire length of the beach below those cliffs in the mid-nineteenth century and wrote of his travels in *Cape Cod,* about 6 km^2 of coastal land has been eaten by the ocean, equivalent to about 150 m of beach retreat.

It has been estimated that more than 70 percent of the total length of the world's sand beaches has retreated in recent decades, at a rate of at least 10 cm per year, and 20 percent of the total length has retreated at a rate of more than 1 m per year. Much of this movement may be traced to the damming of rivers, which decreases the sediment supply to the shoreline.

Depositional Coastal Forms Sediment builds up in areas where tectonic subsidence depresses the crust along a coast. Such coasts are characterized by long, wide beaches and wide, low-lying coastal plains of sedimentary strata. Shoreline forms include sandbars, low-lying sandy islands, and extensive tidal flats. Long beaches grow longer as longshore currents carry sand to the downcurrent end of the beach. There it builds up, first as a submerged bar, then rising above the surface and extending the beach by a narrow addition called a **spit** (Figure 17.17).

Long sandbars offshore may build up and become **barrier islands,** forming a barricade between open ocean waves and the main shoreline (Figure 17.18). Barrier islands are common, especially along low-lying coasts of easily erodible and transportable sediments or poorly cemented sedimentary rocks where longshore currents are strong. Some of the most prominent barrier islands are found along the coast of New Jersey, at Cape Hatteras, and along the Texas coast of the Gulf of Mexico, where one, Padre Island, is 130 km long. As the bars build up above the waves, vegetation takes hold, stabilizing the islands and helping them resist wave erosion during storms. Barrier islands are separated from the coast by tidal flats or shallow lagoons. Like beaches on the main shore, barrier islands are in dynamic equilibrium with the forces shaping them. If their equilibrium is disturbed by natural changes in climate or wave and current regimes, or by real estate development, they may be disrupted or devegetated, leading to increased erosion and even disappearance. Other barrier islands may grow larger and more stable.

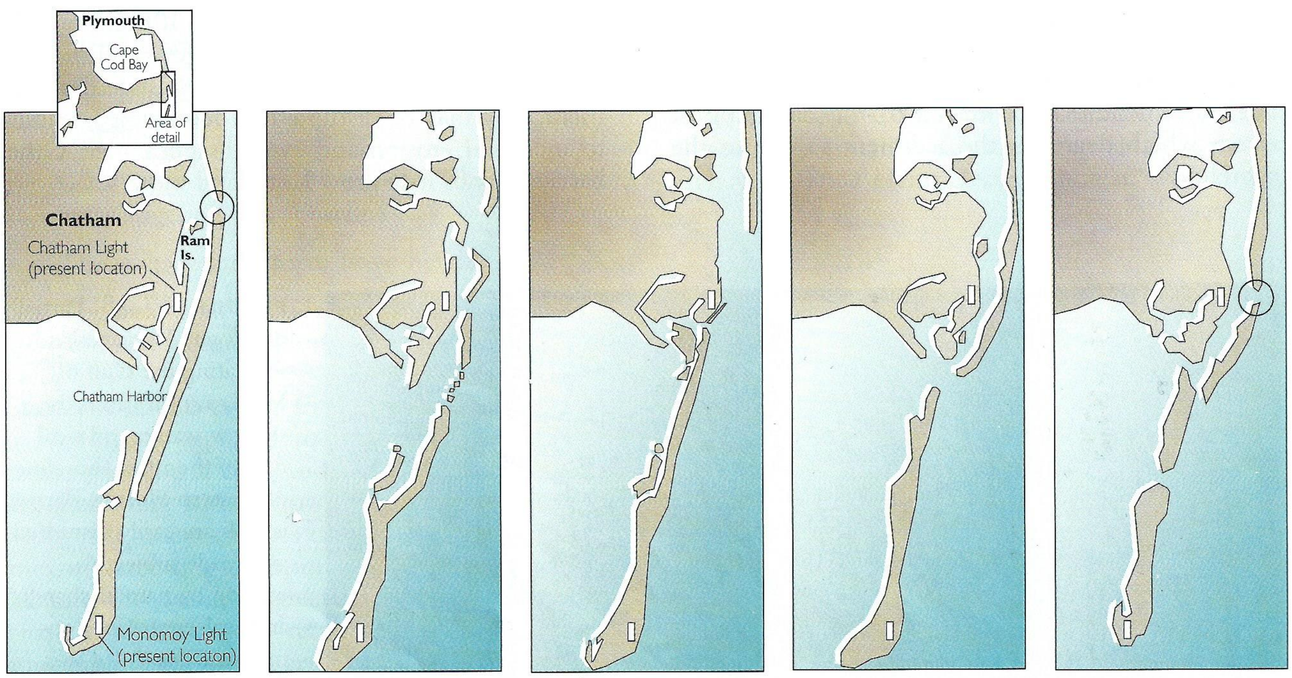
FIGURE 17.18 Partially developed barrier islands separating the Gulf of Mexico *(right)* from the shallow waters and tidal flats of the main shoreline of western Florida. Barrier islands are easily disturbed and made subject to erosion by natural changes and real-estate development. *(Richard A. Davis, Jr.)*

Over the course of hundreds of years, shorelines may undergo significant changes. Hurricanes and other intense storms, such as the "storm of the century" that hit the East Coast in March 1993, may form new inlets or elongate spits or may breach them. Such changes have been documented by remapping at various time intervals from aerial photographs. The shoreline at Chatham, Massachusetts, at the elbow of Cape Cod, has changed enough over the past 160 years that a lighthouse has had to be moved. Figure 17.19 illustrates the numerous changes that have taken place in the configuration of the bars to the north and the long spit of Monomoy Island, as well as several breaches of the bars. Many homes are now at risk in Chatham, but there is little the residents or the state can do to prevent these beach processes from taking their natural course.

Changes in Sea Level

Shorelines are sensitive to changes in sea level, which can change the approach of waves, alter tidal heights, and affect the path of longshore currents. Rise and fall of sea level can be local, a result of tectonic subsidence or uplift, or global, the result, for example, of continental glacial melting or growth (see Chapter 15).

Sea-level changes are detected by geologic studies of wave-cut terraces. These terraces, which were cut at sea level, are now elevated, showing that, at least locally, sea level has dropped in the intervening period (Figure 17.20). A new technique uses satellite altimeter measurements to determine the altitude of the sea surface relative to the carefully determined orbit of the satellite (see Feature 17.2). A study shows that over a two-year period, these measurements can determine differences in the altitude of the sea surface with the remarkable degree of precision of 4 mm. This precision has allowed us to estimate a global sea-level change of about 4 mm per year. Much of this rise may be due to short-term variations, but researchers hope that this method will give us reliable information on rises in sea level caused by global warming. Recent studies indicate that human activities over the globe may be affecting sea level, producing the small rise recorded over the twentieth century. These activities include impoundment of



1830–1850 Circle shows approximate location of 1846 breach in barrier spit. Ram Island later disappears

1870–1890 The beach south of the inlet breaks up and migrates southwest toward the mainland and Monomoy

1910–1930 The southern beach has disappeared, and its remnants soon will connect Monomoy to the mainland

1950–1970 The northern beach steadily grows with cliff sediment; Monomoy breaks from the mainland

1987 140-year cycle begins again with Jan. 2 breach in the barrier spit across from the Chatham Light

FIGURE **17.19** Changes in the shoreline at Chatham, Massachusetts, at the elbow of Cape Cod, over the past 160 years. (After Cindy Daniels, *Boston Globe,* February 23, 1987.)

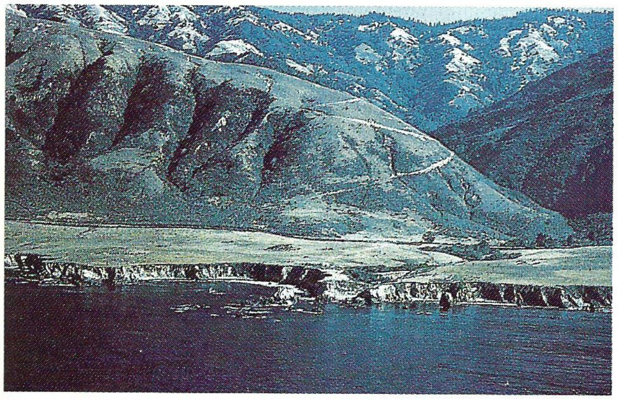


FIGURE 17.20 Raised shoreline terraces caused by coastal emergence indicate a fall in sea level since the terrace cuts were formed at the higher sea level. *(John S. Shelton.)*

surface waters behind dams, draining of wetlands, deforestation, groundwater withdrawal, and surface water diversion. According to one estimate, such activities may have contributed a part of the observed sea-level rise, which may register 0.3 to 1.1 m by the end of the twenty-first century. Others dispute the magnitude of that estimate, but it is clear that sea level may be at least slightly affected by human works, a conclusion that few could have foreseen.

During periods of lowered sea level, areas that were offshore are exposed to agents of erosion. Rivers extend their courses over formerly submerged regions and cut valleys into newly exposed coastal plains. When sea level rises, flooding the lands of the backshore, river valleys are drowned, marine sediments build up along former land areas, and erosion is replaced by sedimentation. Today, long fingers of the sea indent many of the shorelines of the northern and central Atlantic coast. These long indentations are former river valleys that were flooded as the last glacial age ended about 10,000 years ago and the sea level rose.

Drowned river valleys are one kind of **estuary,** which is a coastal body of water connected to the ocean and also supplied with fresh water from a river. The fresh water comes down the river and mixes with seawater in the estuary long before it reaches the main shoreline. At its upper reaches—in some places, many kilometers upstream from the mouth—an estuary is fresh. Downstream, it becomes saltier as it gradually mixes with seawater. By the time it nears the main shoreline, the estuary is entirely seawater.

The shorelines of the world serve as our barometers of impending change as they respond to altered global and local conditions. For example, if global warming causes sea level to rise, we will first see the effects on our beaches. The pollution of our inland waterways sooner or later arrives at our beaches, as sewage from city dumping and oil from ocean tankers wash up on the shore. And as real estate development and construction along shorelines expands, we will see the continuing contraction and even disappearance of some of our finest beaches.

Profound as the geological changes in shorelines may be, they are dwarfed by the geologic processes at work in the vast bulk of the oceans where the water is deep and hides the active seafloor below. In the rest of this chapter, we explore some of the important geologic processes of the deep ocean. We begin with the unusual tools available for measuring and mapping the seafloor.

Sensing the Floor of the Ocean

The best way to see the seafloor is directly from a deep-diving submersible. Pioneered by the French oceanographer Jacques-Yves Cousteau, these small ships can observe and photograph at great depths. With their mechanical arms, they can break off pieces of rock, sample soft sediment, and catch specimens of exotic deep-sea animals. Newer robotic submersibles are guided by scientists on the mother ship above (Figure 17.21). But submersibles are expensive to build and operate, and they cover small areas at best.

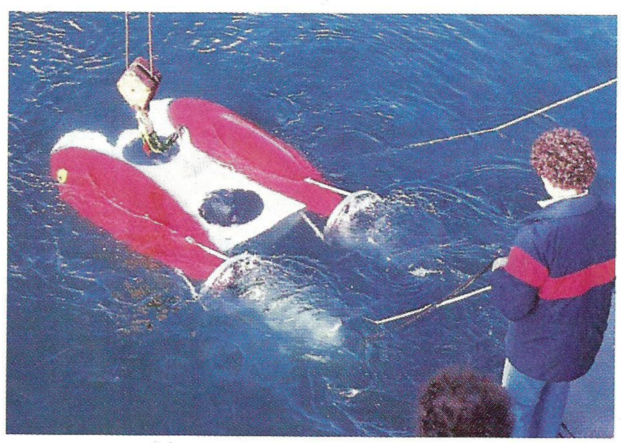


FIGURE 17.21 The *Benthic Explorer.* This small robotic vehicle can explore the seafloor while being directed from shipboard. *(T. Kleindinst/Woods Hole Oceanographic Institution.)*

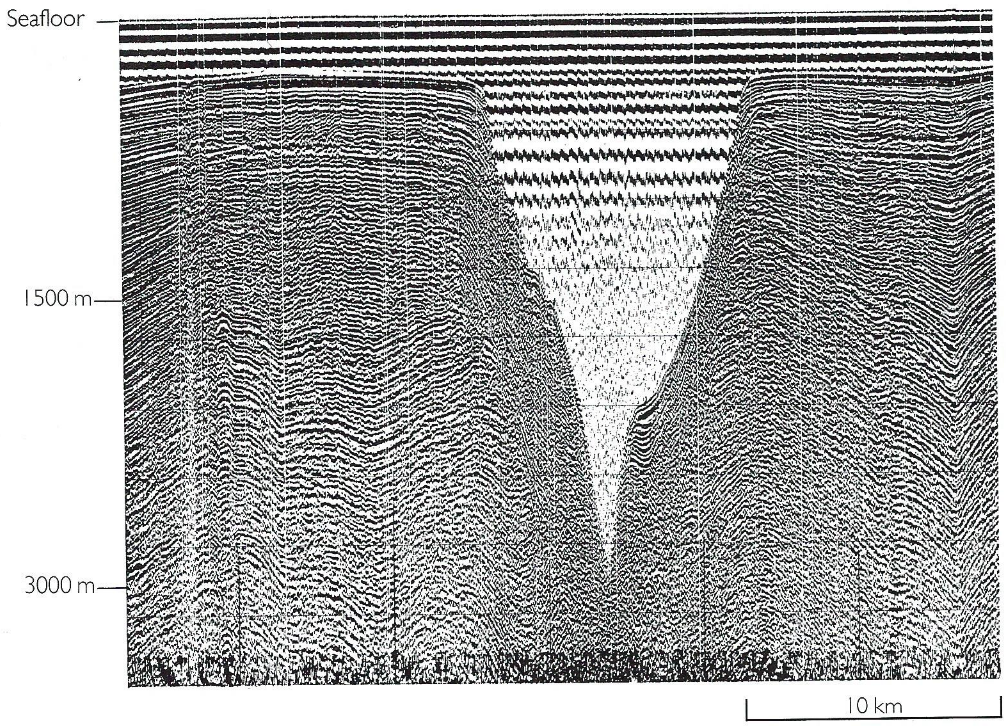


FIGURE 17.22 An echo-sounding profile of the Congo submarine canyon off the Republic of Zaire on the west coast of Africa. The bottom of the canyon at this point is about 3000 m below the seafloor of the continental shelf; at the top, the canyon is more than 10 km wide. The wavy lines below the seafloor surface are sound reflections from bedding planes in continental shelf sediments, somewhat deformed because of mild tectonic disturbance. *(K. O. Emery, Woods Hole Oceanographic Institution.)*

For most work, today's oceanographers use instrumentation to sense the seafloor topography indirectly from a ship at the surface. One shipboard instrument is an echo sounder, which sends out pulses of sound waves. When the sound waves are reflected back from the ocean bottom, they are picked up by sensitive microphones in the water. By measuring the interval between the time the pulse leaves the ship and the time it returns as a reflection, and by figuring in the speed of sound in water, oceanographers can compute the depth. The result is an automatically traced profile of the bottom topography. (Figure 17.22 shows an example.) Echo sounding is also used to probe the stratigraphy of sedimentary layers beneath the ocean floor (see Feature 19.1).

Many other instruments are lowered to the bottom to detect such properties as the magnetism of the seafloor, the shapes of undersea cliffs and mountains, and heat coming from the interior. One of the most spectacular of these instruments are the underwater radar beams used to map the details of topography of the bottom (Figure 17.23). Several different systems use one or more beams that cut varying swaths both ahead and to the side, thus illuminating topography with a detail and accuracy not possible with echo sounders. On a larger scale, underwater cameras can

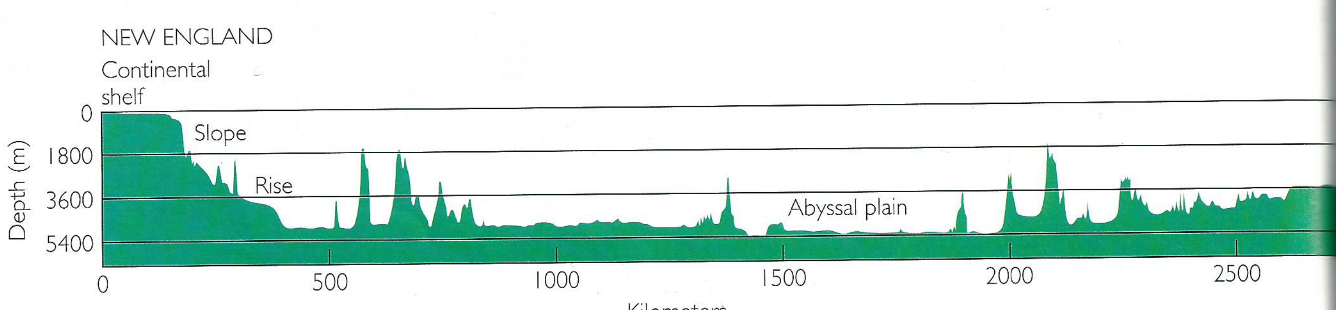


FIGURE 17.24 Topographic profile of the floor of the Atlantic Ocean from New England *(left)* to Gibraltar *(right)*. (After B. C. Heezen, "The Origin of Submarine Canyons," *Scientific American,* August 1956, p. 36.)

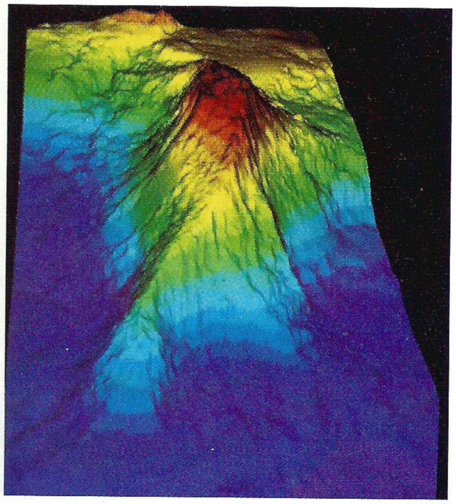


FIGURE 17.23 Loiki Seamount just south of the Big Island of Hawaii, as imaged by computer-enhanced side-scan radar. Loiki is the newest in the string of hot-spot volcanoes that form the Hawaiian island chain. *(Ocean Mapping Development Center, University of Rhode Island.)*

photograph the details of the seafloor surface and the organisms that inhabit the deep. Combined instrument packages that include radar, photography, gravity, and magnetic measurements are towed near the bottom to give even greater accuracy. To be a marine geologist today places one at the center of a beehive of high technology.

Since 1968, the United States–sponsored Deep Sea Drilling Program and its successor, the international Ocean Drilling Program, have sunk hundreds of drill holes to depths of many hundreds of meters below the seafloor. Cores obtained from these drill holes have given us an unprecedented three-dimensional picture of the seafloor and provided samples for detailed physical and chemical studies. (In Chapter 20, we discuss the role of these drilling programs in the study of plate tectonics.)

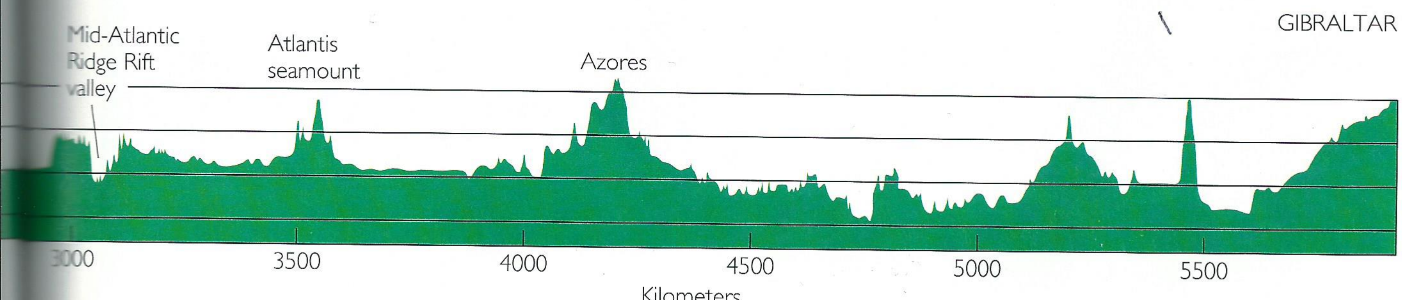
Profiles of Two Oceans

An Atlantic Profile

Just as we profiled North America in Chapter 3 by taking a hypothetical trip across the continent, we can imagine driving a deep-diving submarine along the floor of the Atlantic Ocean from North America to Gibraltar. A topographic profile of the floor of the Atlantic is shown in Figure 17.24.

Starting from the coast of New England, we would descend from the shoreline to depths of 50 to 200 m and travel along the **continental shelf,** a broad, flat, sand- and mud-covered platform that is part of the continent but slightly submerged. After traveling about 50 to 100 km across the shelf, down a very gently inclined surface, we would find ourselves at the edge of the shelf, where we would start down a steeper incline, the **continental slope.** This slope, covered mostly with mud, descends at an angle of about 4°, a drop of 70 m over a horizontal distance of 1 km, which would feel like a noticeable grade if we were driving on land.

The continental slope is irregular and marked by gullies and **submarine canyons,** deep valleys eroded into the slope and the shelf behind it (see Figure 17.22). On the lower parts of the slope, at depths of around 2000 to 3000 m, the incline becomes gentler. Here it merges into a more gradual incline called the **continental rise,** an apron of muddy and sandy sediment extending into the main ocean basin. The rise is broken by an occasional shallow channel or canyon.



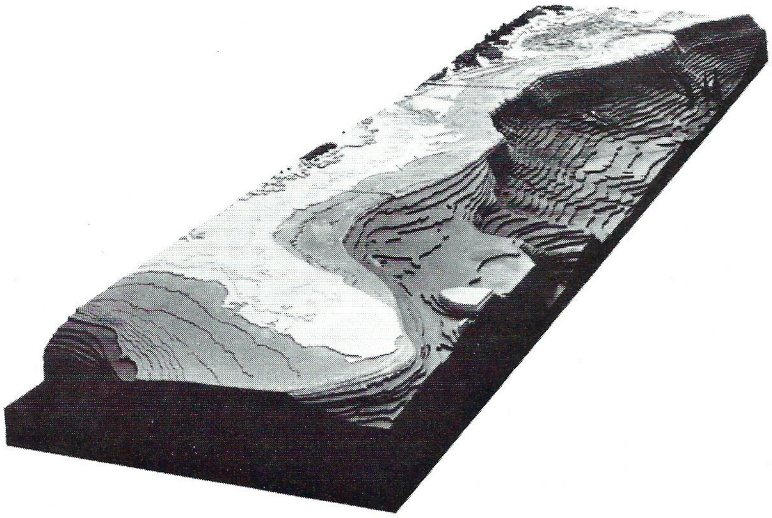


FIGURE 17.25 The Atlantic continental shelf, slope, and rise off the eastern coast of North America.

Figure 17.25 depicts a continental shelf, slope, and rise.

The continental rise is hundreds of kilometers wide, and it grades imperceptibly into a wide, flat **abyssal plain** that covers large areas of the ocean floor at depths of about 4000 m to almost 6000 m (Figure 17.26). These plains are broken by occasional submerged volcanoes, mostly extinct, called **seamounts** (see Figure 17.23). A few seamounts extend to the surface as islands. The Azores are such volcanic islands. In the Caribbean Sea, many of the volcanic islands are capped with coral reefs and other kinds of limestone.

As we travel along the abyssal plain, we gradually climb into a province of low abyssal hills whose slopes are covered with fine sediment. Continuing up the hills, the sediment layer becomes thinner and outcrops of basalt appear beneath it. As we rise along this steep, hilly topography to depths of about 3000 m, we are climbing the flanks and then the mountains of the Mid-Atlantic Ridge.

Abruptly, we come to the edge of a deep, narrow valley (about 1 km wide) at the top of the ridge, a narrow cleft marked by active volcanism. This is a rift valley where two plates separate. As we cross the valley and climb the east side, we are moving from the North American Plate to the Eurasian Plate (see the plate map in the front endpapers).

Continuing east, we find the same topography as on the west side of the ridge, only in reverse order, for the ocean floor is symmetrical on either side of the ridge. Again we pass over abyssal hills; gradually

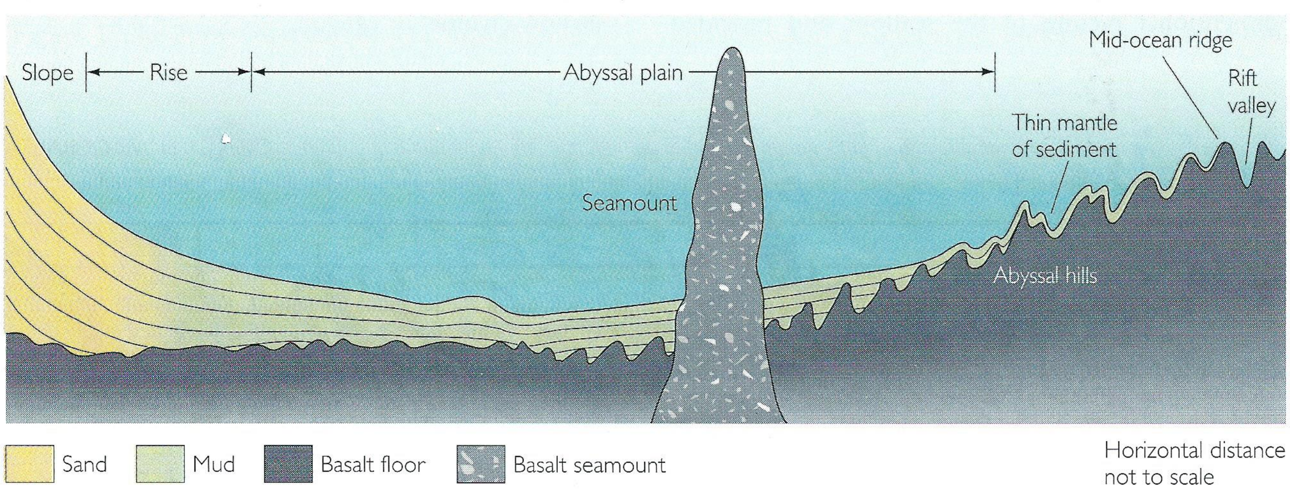


FIGURE 17.26 Profile from continental rise to Mid-Atlantic Ridge.

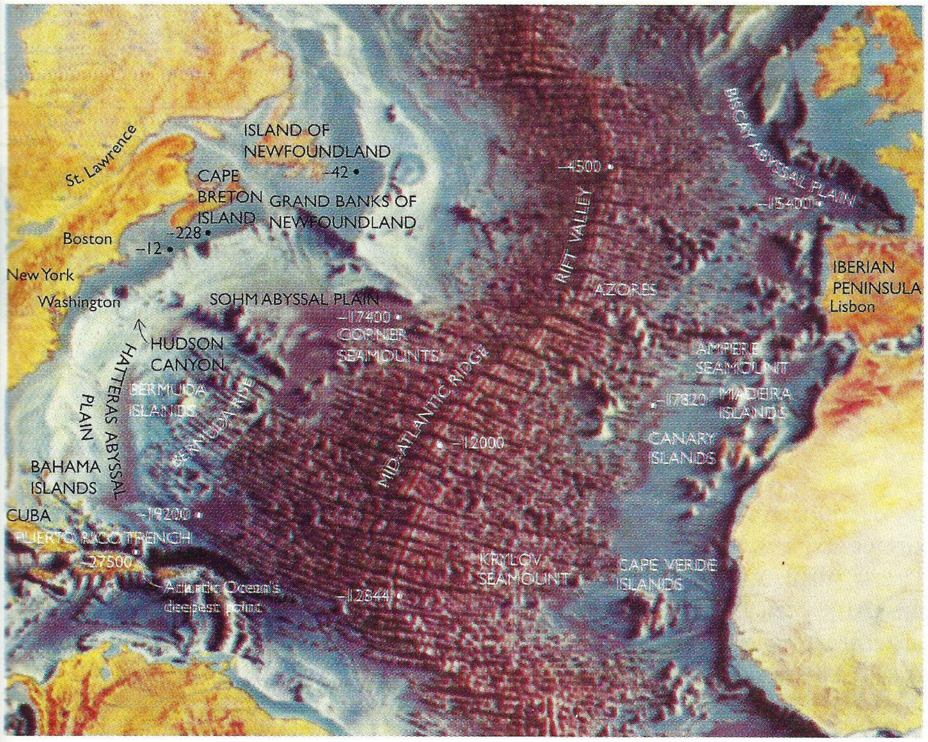


FIGURE 17.27 Artist's representation of the North Atlantic Ocean floor from echo-sounding data. Depths shown are in feet below sea level. (Detail from *World Ocean Floor* by Heinrich C. Berann, based on bathymetric studies by Bruce C. Heezen and Marie Tharp. Copyright © Marie Tharp.)

descend to an abyssal plain; then ascend to the continental rise, slope, and shelf off the coast of Europe.

Oceanographic research vessels made many such trips, crisscrossing the oceans with echo sounders. By the late 1970s, researchers had mapped the floor of the oceans. On the map shown in Figure 17.27, we can see some of the details of individual ocean-floor provinces of the North Atlantic and get an impression of the strange submarine landscape. This map is still somewhat of an approximation, for many of the areas of the seafloor have never been visited by an echo-sounding research ship, and the mappers had to interpolate some of the details between widely separated ship tracks. The next generation of maps would be based on satellite data (Feature 17.2).

Some of the characteristics of oceans are related to plate-tectonic movements. The Atlantic, bisected by the Mid-Atlantic Ridge, is primarily a spreading ocean and has only a small subduction zone in the Caribbean Sea. In contrast, the Pacific, crossed by the East Pacific Rise, shows a large number of subduction zones that are currently narrowing this ocean.

The entire network of ridges, trenches, and transform faults that bound the world's ocean plates can be seen in the map in Feature 17.2. This map was constructed not from echo sounding but from satellite data kept secret by the Navy during the Cold War and only recently made available to all oceanographers. A Navy satellite called Geosat, which repeatedly orbited the Earth, used a radar altimeter to measure the altitude of the sea surface. The bumps and depressions in the sea surface on this map mimic the highs and lows of the seafloor. For example, a seamount that rises 1500 m from the seafloor produces a bump in the sea surface of about 1.5 m. From such data, we can visualize ridges, trenches, plateaus, and abyssal plains that exist where no ship has ever gone. Marine geologists are now busy studying this map and using the data from the satellite to make detailed studies of seafloor topography that we never knew existed.

A Pacific Profile

Just as continents' profiles may differ, a profile of the Pacific Ocean shows features not seen in the Atlantic profile. If we were to travel westward from South America beginning on the west coast of Peru or Chile, we would, as before, cross a continental shelf. This shelf, however, is only a few tens of kilometers wide and more than 100 m deep. At the edge of the shelf, the continental slope is much steeper and extends down to 8000 m as we enter the Peru-Chile

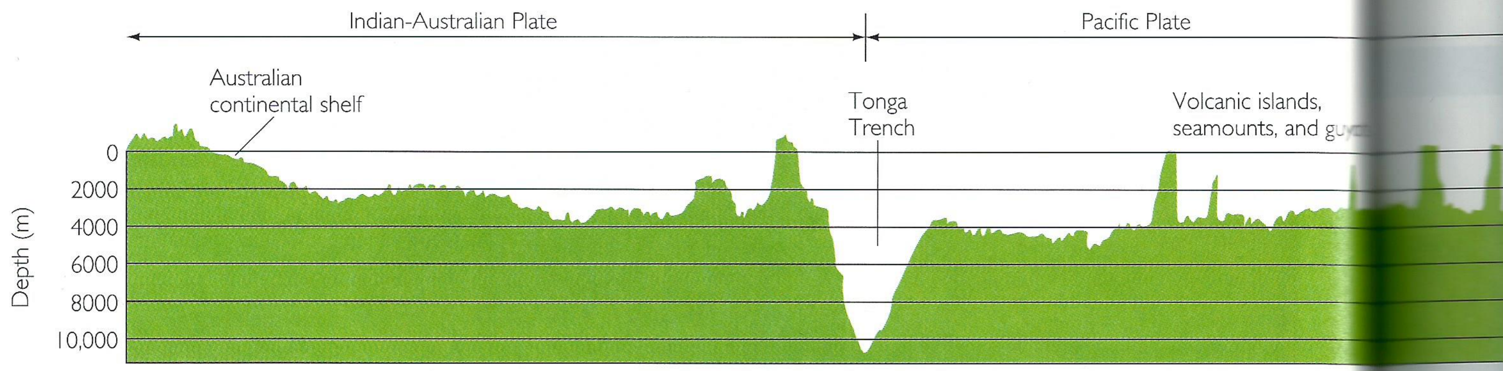


FIGURE 17.28 Topographic profile of the floor of the Pacific Ocean traveling westward from South America *(right)* to Australia *(left)*. (Data synthesized from various sources.)

Trench (Figure 17.28). This long, deep, narrow depression in the seafloor is the surface expression of the subduction of the Nazca Plate, a small plate in the eastern Pacific, under the South American Plate.

Continuing across the trench and up onto the higher hilly region of the Nazca Plate, we soon come to a mid-ocean ridge, the East Pacific Rise. The East Pacific Rise is lower than the Mid-Atlantic Ridge, but it has the characteristic central rift valley and outcrops of basalt. On the west side of the East Pacific Rise, we cross over to the Pacific Plate and drive on westward over its broad central regions. Eventually we come to another trench, the Tonga. This is one of the deepest places in all the oceans, almost 11,000 m deep. Here, in the middle of the ocean, the Pacific Plate subducts beneath the Indian-Australian Plate. On the west side of the trench, an arc of volcanic islands rises from the deep seafloor and erupts basalt and andesite. Leaving the island arc, we return to the deep seafloor, now on the Indian-Australian Plate, and soon come to the continental rise, slope, and shelf of Australia, similar to the east coast of North America.

Continental Margins

The shorelines, shelves, and slopes of the continents are together called **continental margins.** The profiles of the oceans show that there are two types of continental margins: passive and active. A **passive margin** is a continental borderland far from a plate boundary. A good example is the broad region, associated with a spreading ocean, off the east coast of North America (Figure 17.29). Such margins are called passive (implying quiescence) because volcanoes are absent and earthquakes are few and far between. In contrast, **active margins** are associated with subduction zones and transform faults. The volcanic activity and frequent earthquakes give these narrow and tectonically deformed continental margins their name. One active margin at a subduction zone is off the west coast of South America (see Figure 17.28). This margin includes the trench offshore, the narrow shelf, and the active volcanic belt of the Andes Mountains. As the subducting Nazca Plate bends below the continent, it heats up, and magmas form and work their way to the surface.

The continental shelves of passive margins consist of essentially flat-lying, shallow-water sediments, both terrigenous and carbonate, several kilometers thick (see Figure 17.29). Although the same kinds of sediment are found on active margin shelves, they

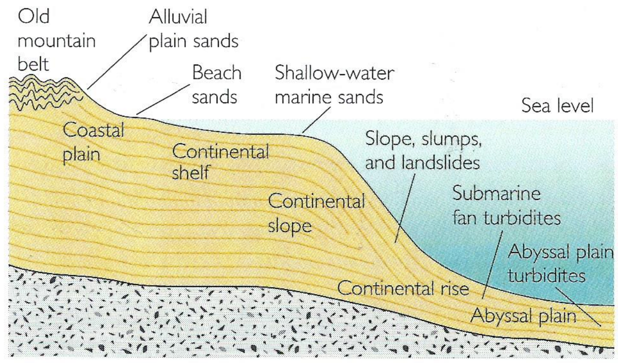
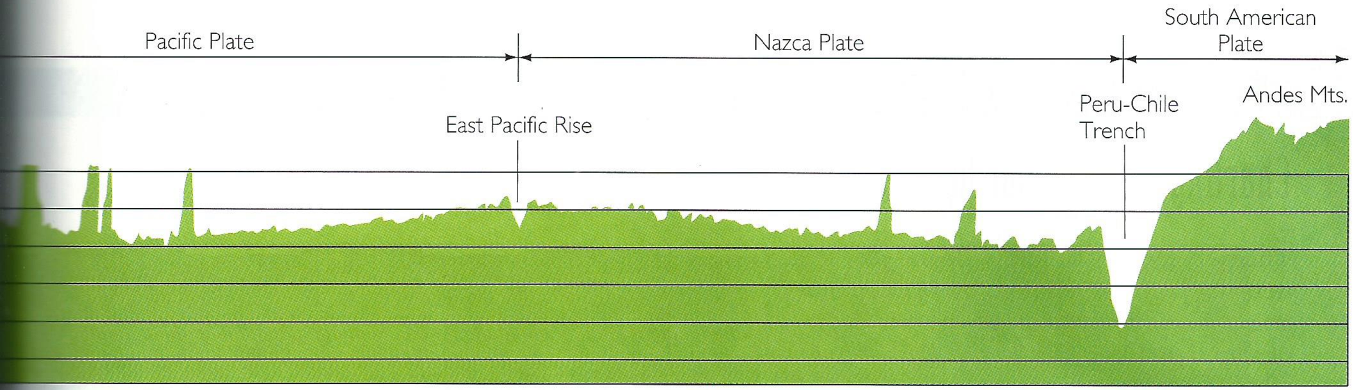


FIGURE 17.29 A profile of the Atlantic passive continental margin off southern New England. (After K. O. Emery and E. Uchupi, *Atlantic Continental Margin of North America,* American Association of Petroleum Geologists, 1972.)



are more likely to be structurally deformed and to include ash and other volcanic materials.

Continental Shelf

The continental shelf is one of the most economically valuable parts of the ocean. Georges Bank off New England and the Grand Banks of Newfoundland, for example, have been among the world's most productive fishing grounds for all of this century. In recent years, the continental shelf, especially off the Gulf coast of Louisiana and Texas, has housed huge oil-drilling platforms. These are some of the reasons why in 1982 most of the world's nations (although not the United States) signed the international Law of the Sea treaty governing territorial and economic rights of nations.

Continental shelves are broad and relatively flat at passive continental margins and are narrow and uneven at active margins. As we noted earlier, because continental shelves lie at shallow depths, they are subject to exposure and submergence as a result of changes in sea level. During the Pleistocene glaciation, all of the shelves now at depths of less than 100 m were above sea level, and many of their features were formed then. At that time, shelves at high latitudes were glaciated, producing an irregular topography of shallow valleys, basins, and ridges. The shelves in lower latitudes were left more regular, broken by occasional stream valleys.

Continental Slope and Rise: Turbidity Currents

The waters of the continental slope and rise are too deep for the seafloor to be affected by waves and tidal currents. As a consequence, muds, silts, and sands that have been carried across the shallow continental shelf come to rest as they are draped over the slope. The slope shows signs of the slumping of sediment and the erosional scars of gullies and submarine canyons. Deposits of sands, silts, and muds on both slope and rise indicate active sediment transport in these deep waters. For some time, geologists were puzzled over what kind of current might cause both erosion and sedimentation on the slope and rise at such great depths.

The answer proved to be a **turbidity current**—a flow of turbid, muddy water down a slope. Because of its suspended load of mud, the turbid water is denser than the overlying clear water and flows beneath it. Turbidity currents were first noticed more than a century ago where the Rhone River enters Lake Geneva in Switzerland. The muddy river water enters the clear water of the lake and flows as a distinct current along the sloping bottom to the level floor of the lake, where it fans out over the bottom.

Turbidity currents can both erode and transport sediment, and their role in ocean processes was first understood by Philip Kuenen, a Dutch geologist and oceanographer. In 1936, Kuenen produced and filmed such currents in his laboratory by pouring muddy water into the end of a long, narrow tank with a sloping bottom. He showed that these currents could move at many kilometers per hour and that the speed was proportional to the steepness of the slope and the density of the current. Kuenen reasoned that because of its speed and turbulence, a turbidity current could erode and transport large quantities of sand down the continental slope. He then proposed the idea, revolutionary for that time, that turbidity currents operate widely in the ocean, especially on continental slopes, at depths well below any possible wave or tidal action.

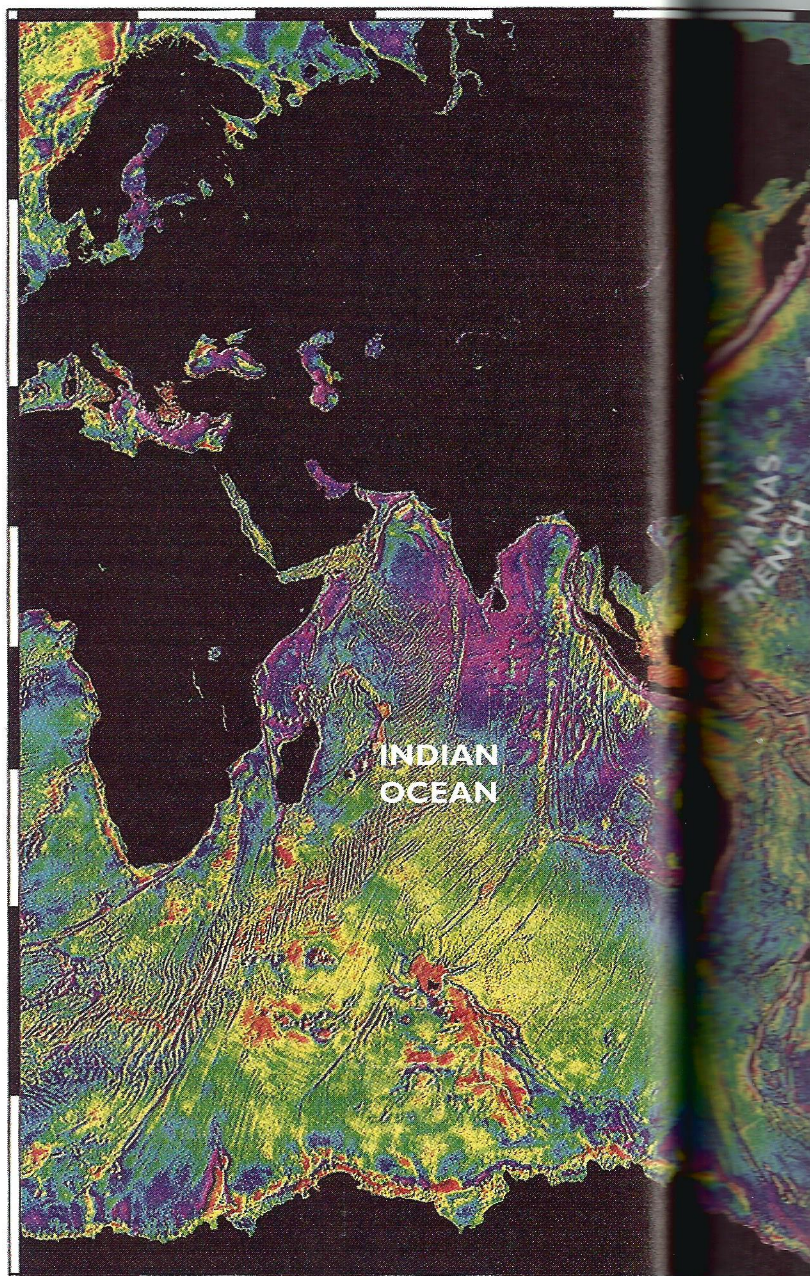
17.2 TECHNOLOGY AND EARTH

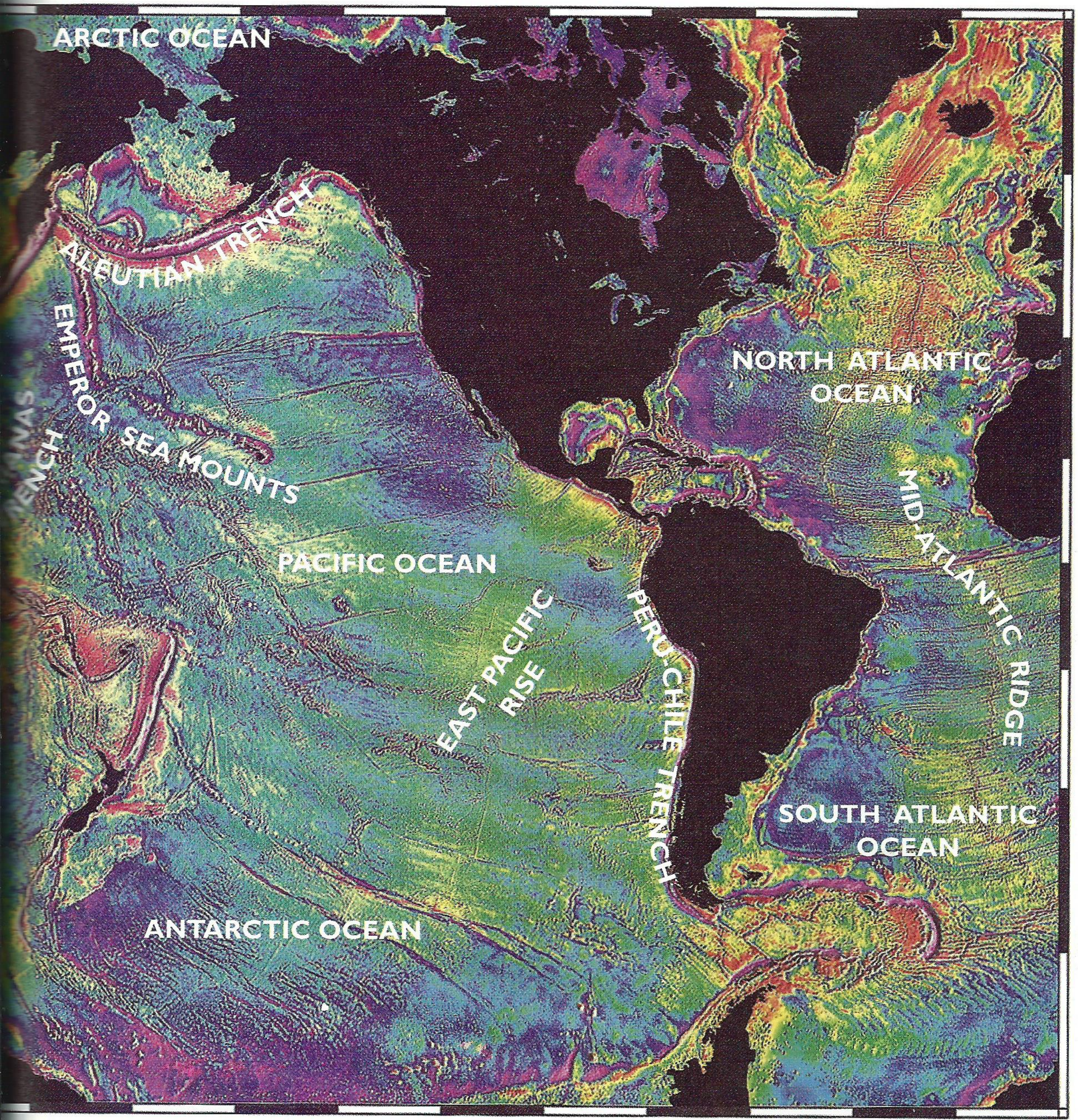
Charting the Seafloor by Satellite

The rich geology of the seafloor—its ridges, trenches, seamounts, and transform boundaries—became apparently only after decades of ship soundings. Our knowledge of regions where few ships travel remains fragmentary. Recently, however, scientists have developed a tool that enables a satellite to "see through" the ocean and chart the topography of the seafloor, gathering data in mere months.

The new method makes use of an altimeter mounted on a satellite. The altimeter sends pulses of radar beams that are reflected back from the ocean below, giving measurements of the distance between the satellite and the sea surface with a precision of a few centimeters. The height of the sea surface depends not only on waves and ocean currents but also on changes in gravity caused by the topography and composition of the underlying seafloor. The gravitational attraction of a seamount, for example, can cause water to "pile up" above it, producing a bulge in the sea surface as much as 5 m above average sea level. Similarly, the diminished gravity over a deep-sea trench would show as a depression of the sea surface of as much as 60 m. In this way, features of the ocean floor can be inferred from satellite data and displayed as if the seas were drained away.

In the satellite photograph shown here, shallow regions, deep regions, and intermediate depths can be distinguished. Also clearly visible are the raised stripe between Europe and North America that marks the Mid-Atlantic Ridge and its associated transform boundaries, the trail of a hot spot in the Pacific marked by the Emperor-Hawaiian seamount chain, and the major deep-sea trenches at subduction boundaries. New features not revealed by ship surveys have already been found, and future surveys may reveal even deeper structures, such as convection currents in the mantle.





Marine gravity anomaly from satellite altimetry. *(D. T. Sandwell and W. H. F. Smith/Geological Data Center, Scripps Institution of Oceanography, University of California, San Diego.)*

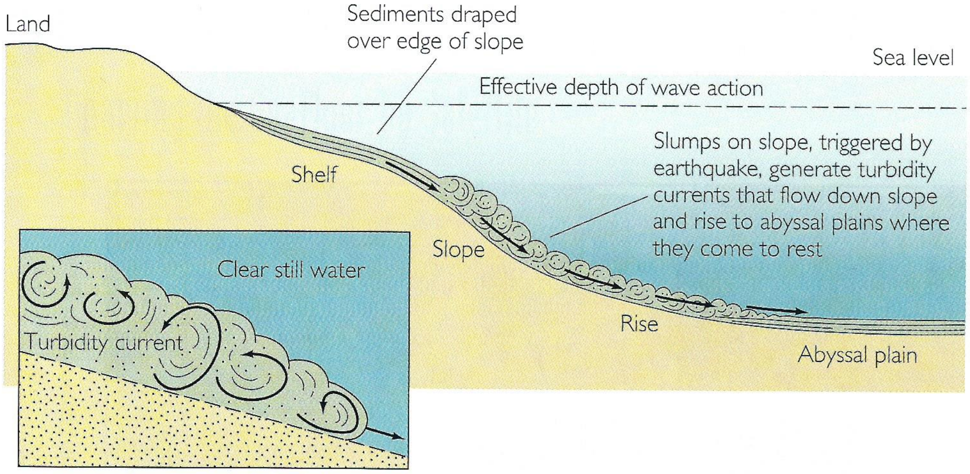
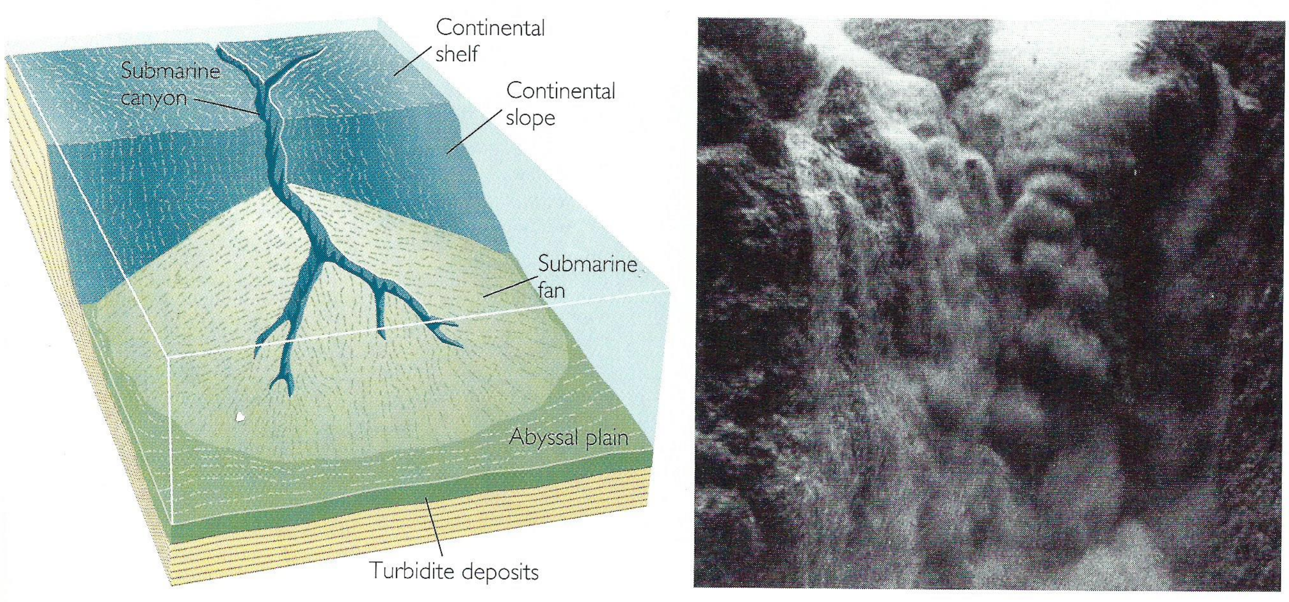


FIGURE 17.30 How a turbidity current forms in the ocean. These currents can erode and transport large quantities of sand down the continental slope.

Turbidity currents start when occasional earthquakes trigger slumps of the sediment draped over the edge of the continental shelf and onto the continental slope (Figure 17.30). The sudden slump, or submarine landslide, throws mud into suspension, creating a dense, turbid layer of water near the bottom. This turbid layer starts to flow, accelerating down the slope.

As the turbidity current reaches the foot of the slope and the gentler incline of the continental rise, the current slows and some of the coarser sandy sediment starts to settle, often forming a **submarine fan,** a deposit something like an alluvial fan on land. Many currents continue across the rise, cutting channels in the submarine fans (Figure 17.31). Eventually, the currents reach the level bottom of the ocean basin, the abyssal plain, where they spread out and come to rest in graded beds of sand, silt, and mud called **turbidites.**



(a) (b)

FIGURE 17.31 (a) Submarine canyons and fans are formed by turbidity currents that start on the continental shelf or slope and erode canyons in them, leading to channels on the fan. (b) Sandfall at the head of a submarine canyon at the edge of the continental shelf. These falls generate sandy flows, like turbidity currents, that lay down fans of sandy sediment at the foot of the continental slope. *(U.S. Navy.)*

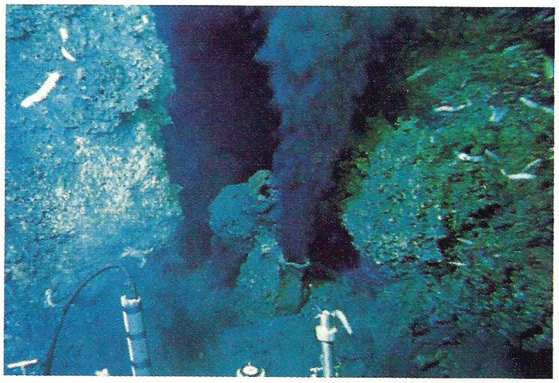
17.3 INTERPRETING THE EARTH

Hot Springs on the Seafloor

Much of the heat coming from Earth's interior is thought to be dissipated when cold seawater percolates into the many fissures associated with seafloor spreading along mid-ocean ridges. The cold water sinks several kilometers, encounters hot basalt, surges upward, and emerges on the seafloor as hot springs, rich in dissolved minerals and gases leached from the magma. This hypothesis was first verified in 1977; subsequently, spectacular discoveries of such hydrothermal vents at many places along different mid-ocean ridges have been reported. The Galápagos Islands rift zone spreading center was the first to be discovered, followed by reports of vents at several other places on the East Pacific Rise, on the Gorda and Juan de Fuca ridges in the Pacific off the shores of Oregon and Washington, and at several places along the Mid-Atlantic Ridge. Vents have also been explored in the western Pacific, at the Mariana Trough, and in several other locations. It is now certain that vents are widely distributed over the oceans, especially along mid-ocean ridges.

The hot springs seem to take two main forms. In the Galápagos Islands rift zone, the springs flow gently from cracks at maximum temperatures of about 16°C. On the East Pacific Rise, near Baja, California, superheated water (380°C) spouts forcefully from mineralized chimneys. The chimneys are built up from the dissolved minerals that precipitate around the hot jet as it mixes with the near-freezing waters on the ocean bottom. The deep-diving submarine *Alvin,* not built for such high temperatures, was nearly destroyed when it first approached a superheated vent.

Hydrothermal vents represent a major ore-forming process, possibly an important new source of such minerals as sulfide ores rich in zinc, copper, and iron. Once every 8 million years, the entire ocean cycles through such hydrothermal systems, and this process profoundly affects ocean chemistry by transporting elements from the interior to the ocean. The ecology of these vents is completely different from that of the dark, near-freezing, barren ocean bottom at great depths. Dense colonies of exotic life forms populate the warm water surrounding the vents. Among them are new species of giant worms, clams, and crabs. The vent animals live on unusual primitive bacteria called the *Archaea* that draw energy from the hydrogen sulfide, carbon dioxide, and oxygen in the vent water rather than from the Sun, as the species at the ocean surface do. The Archaea have been discovered to be a major group of organisms, on a parallel with bacteria and multicellular life forms. A new chapter of ocean science was opened by the deep-sea explorers who discovered the hot springs of the seafloor and verified the dominant role of circulating seawater in cooling the new crust formed at ocean ridges.



A plume of hot, mineral-laden water spouts from a hydrothermal vent on the East Pacific Rise. *(D. B. Foster/Woods Hole Oceanographic Institution.)*

For more than a decade after Kuenen's findings, some scientists challenged the existence of turbidity currents, but dramatic confirmation of the hypothesis came from an unexpected source. Transatlantic telegraph cables that had been laid on the ocean floor from North America to Europe were known to break periodically. In 1929, following an earthquake, a particularly large series of cable breaks was reported on the Atlantic continental slope and rise off the Grand Banks of Newfoundland.

In 1952, oceanographers Bruce Heezen and Maurice Ewing of Columbia University, impressed with Kuenen's work, plotted the exact times and positions of these breaks. Could slope slumps and turbidity currents explain the breaks? A rapid breaking of the cables high on the slope was followed by a sequence of breaks going down the slope, farther and farther from the center of the earthquake. If the breaks downslope had been caused by earthquake waves, they should have occurred much earlier. The only reasonable explanation was that the earthquake triggered a slump, which activated a turbidity current fast and powerful enough to snap the cables as it raced down the slope and rise. Later, graded beds were recovered from the seafloor in the path of the presumed flow. As researchers plotted similar patterns of cable breaks in other places, they realized that the turbidity current theory had been confirmed.

Today, we recognize turbidity currents as important agents of erosion and sedimentation on continental margins. They are found in trenches along active margins as well as on the continental slopes and rises of passive margins. They form the abyssal plains that cover large areas of the ocean floor.

Submarine Canyons

Submarine canyons are deep valleys eroded into the continental shelf and slope. They were discovered near the beginning of the twentieth century and were first mapped in detail in 1937. Almost immediately after ocean turbidity currents were hypothesized, they were proposed to be the erosive agents that incised submarine canyons into many continental shelves (see Figure 17.31). This idea, like the currents themselves, remained controversial for many years. Even though submarine canyons have been mapped in detail and their walls and floors amply photographed and sampled, they have been among the most perplexing topographic features of the seafloor. When the canyons were first discovered, some geologists thought they might have been formed by rivers. But this hypothesis soon proved impossible as the complete explanation. Most of the canyon floors are thousands of meters deep, far below the approximately 100-m depth to which rivers could erode during the maximum lowering of sea level in the ice ages. Even so, there is no question that the shallower parts of some canyons were river channels during periods of low sea level.

Although other types of currents have been proposed, turbidity currents are now the favored explanation for the deeper parts of submarine canyons. Evidence supporting this conclusion comes in part from a comparison of modern canyons and their deposits with well-preserved similar deposits of the past, particularly the pattern of turbidites deposited on submarine fans.

The Floor of the Deep Ocean

The deep seafloor is constructed primarily by volcanism related to plate-tectonic motions and secondarily by sedimentation in the open sea. When plates grow by spreading from a mid-ocean ridge, huge quantities of basalt well up from the mantle, forming the oceanic lithosphere. As the plates spread away from the ridge, the basaltic lithosphere cools and contracts, lowering the seafloor. While this is happening, the basalt surface receives a steady rain of sediment from surface waters and gradually becomes mantled with deep-sea muds and other deposits.

Mid-Ocean Ridges

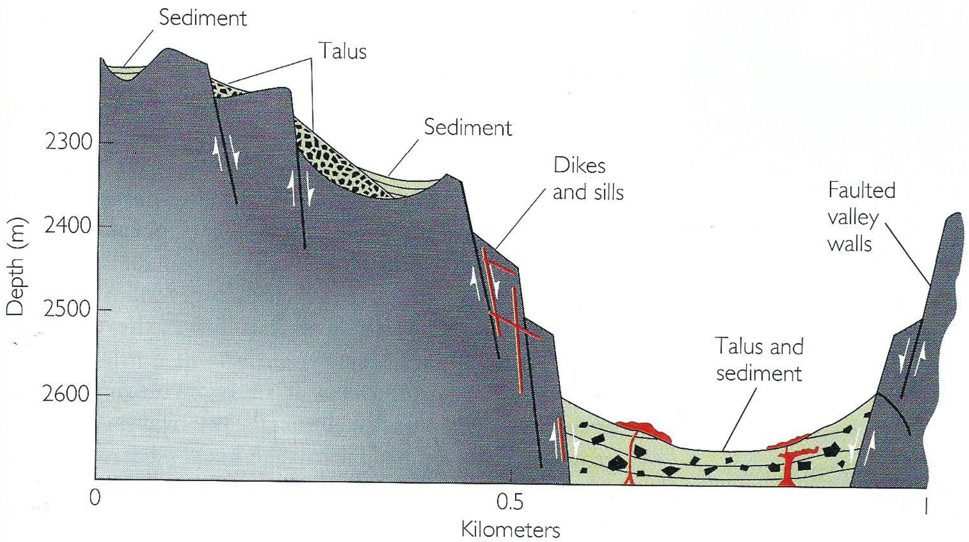
Mid-ocean ridges are the sites of the most intense volcanic and tectonic activity on the deep seafloor. The main rift valley is the center of the action. The valley walls are faulted and intruded with basalt sills and dikes (Figure 17.32), and the floor of the valley is covered with flows of basalt and talus blocks from the valley walls, mixed with a little sediment settling from surface waters. Mid-ocean ridges are offset at many places by transform faults that laterally displace the rift valleys (see Figure 17.27).

Hydrothermal springs form on the rift valley floor as seawater percolates into cracks and fractures in the basalt on the flanks of the ridge, is heated as it moves down to hotter basalt, and finally exits at the valley floor, where it boils up at temperatures as high as 380°C (see Feature 17.3). Some springs are "black smokers," full of dissolved hydrogen sulfide and metals leached from the basalt by the hot waters. Others are "white smokers" with a different composition and lower temperatures. Hydrothermal springs on the seafloor produce mounds of iron-rich clay minerals, iron and manganese oxides, and large deposits of iron-zinc-copper sulfides.

Hills and Plateaus

The floor of the deep oceans away from mid-ocean ridges is a landscape of hills, plateaus, sediment-floored basins, and seamounts. Most of the thousands of volcanoes are submerged, but some rise to the sea surface. Seamounts and volcanic islands may be isolated, in clusters, or in chains. They may be formed along a mid-ocean ridge or where a plate overrides a mantle hot spot. Many seamounts have flat tops, the result of erosion of an island volcano when it was above sea level. These **guyots,** as they are called, are submerged because the plate they were riding on cooled, contracted, and subsided as it passed away from the hot spot that produced the upwelling basalt from the mantle.

Abyssal hills, plateaus, and low ridges are all accumulations of volcanic rock. Many of these features form when the seafloor first opens at a rift valley. Others form as volcanic chains are created over hot spots. The traces of transform-fault offsets of the mid-ocean ridges interrupt the abyssal seafloor with long ridges coupled with parallel valleys, both roughly at right angles to the ridge (see Figure 17.27 and Feature 17.2). One of the largest plateaus, and probably the oldest (Jurassic-Cretaceous boundary), is the Shatsky Rise in the northwest Pacific Ocean, about 1600 km southeast of Japan. The Shatsky Rise is thought to have accumulated from the outpouring of basalt at a former hot spot, the product of a very large mantle plume or set of plumes.



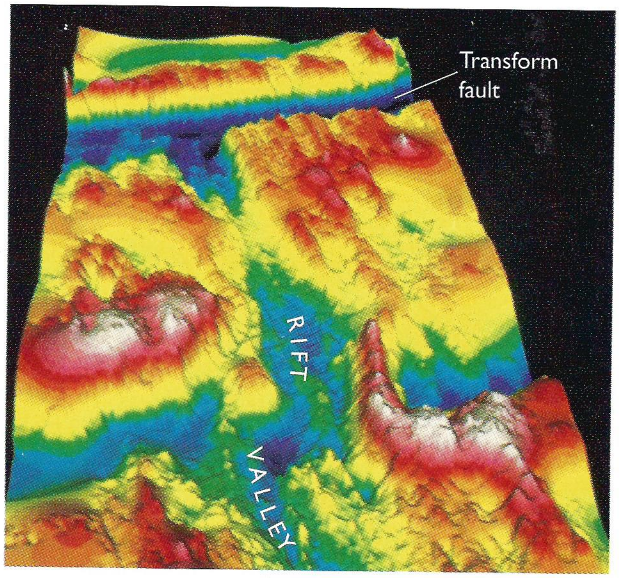


FIGURE 17.32 *Above:* A profile of the central rift valley of the Mid-Atlantic Ridge in the FAMOUS (French-American Mid-Ocean Undersea Study) area southwest of the Azores Islands. The deep valley, where most of the basalt is extruded, is faulted. (After ARCYANA, "Transform Fault and Rift Valley from Bathyscaph and Diving Saucer," *Science,* vol. 190, 1975, p. 108.) *Right:* Computer-generated image (based on data from multibeam sidescan sonar) of the Mid-Atlantic Ridge at about 31° south latitude reveals a 19-km-wide central valley studded with volcanic cones. The valley is bounded by mountains about 2000 m high. A transform fault intersects the valley at the top of the image. (K. C. Macdonald and P. J. Fox, "The Mid-Ocean Ridge," *Scientific American,* vol. 262, 1990, pp. 72–79.)

FIGURE 17.33 Some of the Maldive Islands, coral atolls of the Indian Ocean southwest of the tip of southern India, each surrounded by a circular coral reef that protects the lagoons. In the background is a reef without a central island. *(Guido Alberto Rossi/The Image Bank.)*

Although the deep seafloor is tectonically active at mid-ocean ridges, subduction zones, and hot spots, it is quiescent over much of the vast areas of abyssal hills and abyssal plains. In these areas, the only geological activity is the slow rain of sediment from surface waters, a feature that has led to serious consideration of some parts of the deep seafloor as a waste repository (see Feature 17.4).

Coral Reefs and Atolls

For more than 200 years, coral reefs have attracted explorers and travel writers. Ever since Charles Darwin sailed the oceans on the *Beagle* from 1831 to 1836, these reefs have been a matter of scientific discussion, too. Darwin was one of the first to analyze the geology of coral reefs, and his theory of their origin is still accepted today.

The coral reefs Darwin studied were **atolls,** islands in the open ocean with circular lagoons enclosed within a more or less circular chain of islands (Figure 17.33). Coral reefs also form at continental margins, such as those of the Florida Keys. The outermost part of a reef is a slightly submerged, wave-resistant reef front, a steep slope facing the ocean. The reef front is composed of the interlaced skeletons of actively growing coral and calcareous algae, forming a tough, hard limestone. Behind the reef front is a flat platform extending into a shallow lagoon. An island may lie at the center of the lagoon. Parts of the reef, as well as a central island, are above water and may become forested. A great many plant and animal species inhabit the reef and the lagoon.

Coral reefs are generally limited to waters less than about 20 m deep, for below that depth seawater does not transmit enough light to enable reef-building corals to grow. (Exceptions are some kinds of individual—noncolonial—corals that grow in much deeper waters.) Darwin explained how coral reefs could be built up from the bottom of the dark, deep ocean. The process starts with a volcano building up to the surface from the seafloor (Figure 17.34). As the volcano temporarily or permanently becomes dormant, coral and algae colonize the shore and build **fringing reefs,** coral reefs similar to atolls that grow around the edges of a central volcanic island. Erosion may then lower the volcanic island almost to sea level.

Darwin reasoned that if such a volcanic island were to slowly subside beneath the waves, actively growing coral and algae might keep pace with the subsidence, continuously building up the reef so that the island remained. In this way, the volcanic island would disappear and we would be left with an atoll. More than 100 years after Darwin proposed his theory, deep drilling on several atolls penetrated volcanic rock below the coralline limestone and confirmed the theory. And some decades later, the theory of plate tectonics explained both volcanism and the subsidence that resulted from plate cooling and contraction.

Physical and Chemical Sedimentation in the Ocean

Almost everywhere that oceanographers search the seafloor, they find a blanket of sediment. The muds and sands mantle and cover the topography of basalt originally formed at mid-ocean ridges. The ceaseless sedimentation in the world's oceans modifies the structures formed by plate tectonics and creates its own topography at sites of rapid deposition. The sediment is mainly of two kinds: terrigenous muds and sands eroded from the continents, and biochemically precipitated shells of organisms that live in the sea. In parts of the ocean near subduction zones, sediments derived from volcanic ash and lava flows are abundant. In tropical arms of the sea where evaporation is intense, evaporite sediments are deposited.

Sedimentation on Continental Margins

Terrigenous sedimentation on the continental shelf is produced by the same forces that form beaches: waves and tides. The waves of large storms and hurricanes move sediment over the shallow and moderate depths of the shelf, and tidal currents also flow over the shelf. The waves and currents distribute the sediment brought in by rivers into long ribbons of sand and layers of silt and mud.

Biochemical sedimentation on the shelf results from the buildup of layers of the calcium carbonate shells of clams, oysters, and many other organisms living in shallow waters. Most of these organisms cannot tolerate muddy waters and are found only where terrigenous materials are minor or absent, such as along the extreme southern coast of Florida or off the coast of Yucatan in Mexico. Here, coral reefs thrive and organisms build up large thicknesses of carbonate sediment.

Deep-Sea Sedimentation

Far from the continental margins, fine-grained terrigenous and biochemically precipitated particles suspended in seawater slowly settle from the surface to the bottom. These open-ocean sediments, called **pelagic sediments,** are characterized by great distance from continental margins, fine particle size, and a slow settling mode of deposition. The terrigenous materials are brownish and grayish clays, which accumulate on the seafloor at a very slow rate, a few millimeters every 1000 years. A small fraction, about 10 percent, may be blown by the wind to the open ocean. Winds from the Sahara Desert, for example,

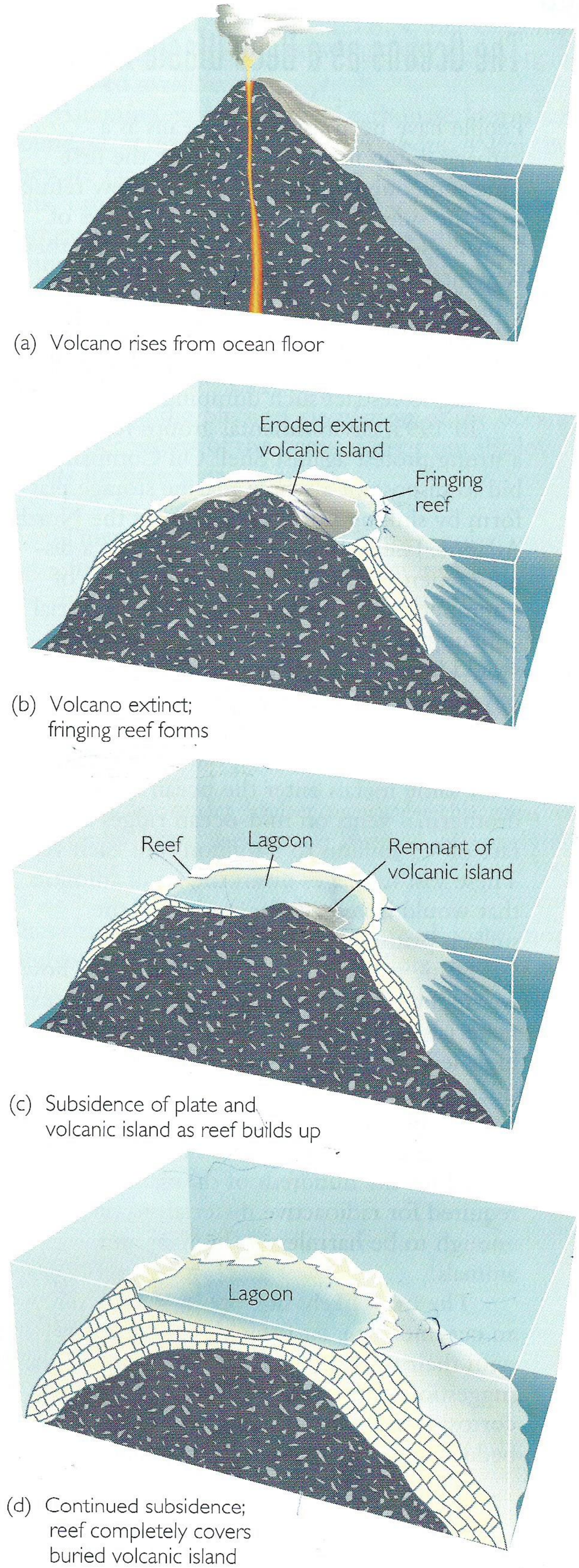


FIGURE 17.34 Evolution of a coral reef from a subsiding volcanic island, first proposed by Charles Darwin in the nineteenth century.

17.4 LIVING ON EARTH

The Oceans as a Deep Waste Repository

People have been using the oceans as a garbage dump for millennia, since the first time someone in a dugout canoe threw refuse over the side. Today, enormous quantities of garbage, sewage, and industrial waste find their way into the sea from ships and coastal communities. Only now, as hypodermic syringes and other medical wastes as well as all sorts of other refuse wash ashore, are some nations beginning to restrict such dumping.

In 1995, environmental groups registered a strong protest against Shell Oil Company's bid to dispose of a whole marine storage platform by sinking it to the bottom of the North Atlantic. The bid became the center of a dispute when it was held up as an example of improper disposal on the seafloor of material containing toxic metals. The waste platform would have included "hundreds of kilograms" of cadmium, mercury, zinc, lead, and nickel. Shell eventually dropped the plan, but some marine geochemists have pointed out that these same metals enter the ocean at hydrothermal vents on mid-ocean ridges at a rate of 0.5 million to 5 million tons each year. These vast tonnages dwarf the small amounts that would have been leached from the sunken platform.

Some oceanographers are evaluating how we might safely dump a variety of hazardous materials, from highly toxic sewage sludges to high-level radioactive wastes, which are considered candidates for burial in the deep ocean. The goal is to deposit the waste in some location where it could remain undisturbed for the hundreds of thousands of years required for radioactive materials to decay enough to be harmless to humans and animals.

The most likely sites are in the middle of an oceanic plate, far from plate boundaries with their tectonic and volcanic activity. One suggestion is to design strong drums of corrosion-resistant material that would be embedded many meters below the seafloor in the accumulated sediment. Proponents of this idea believe the sediment would form a seal around the drums and retard or prevent any leakage that might occur after long burial. For this idea to work economically and safely, we must be able to convert the waste material to a form, such as a dense glass or ceramic, that could be loaded into the drums and would resist chemical attack by seawater or sediment pore waters.

Critics of deep-ocean burial point out that the dangers to ocean life, even at depths where few organisms interact with the bulk of the ocean's populations, are incalculable should the drums prove susceptible to leakage. Their arguments have gained support from the recent revelation that Russian ships have been dumping low-level radioactive waste in the Sea of Japan since the mid-1950s and still continue to do so. Such a violation of a prudent dumping policy has made the public chary of allowing any seabed disposal at all. No agreement has been reached, and the argument continues because waste disposal on land presents different but equally valid hazards.

Plastic debris dumped at sea has washed up onto this beach on the Gulf of Mexico. *(Robert Visser, Greenpeace.)*

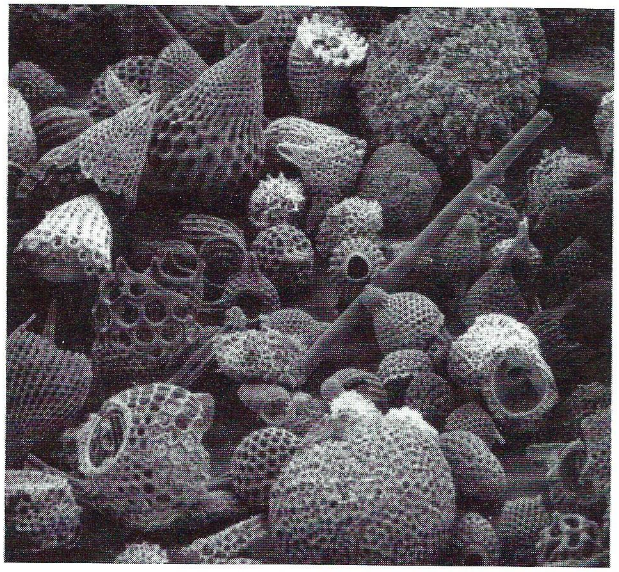


FIGURE 17.35 Scanning electron micrograph of oceanic ooze. Shown here are shells of both carbonate- and silica-secreting unicellular organisms. *(Scripps Institution of Oceanography, University of California, San Diego.)*

blow much dust, silt, and fine sand into the eastern Atlantic off the coast of Africa. Windblown volcanic ash may be deposited downwind from subduction-zone volcanoes.

Within the pelagic sediments, the most abundant biochemically precipitated particles are the shells of foraminifera, tiny single-celled animals that float in the surface waters of the sea. These calcium carbonate shells fall to the bottom after their occupants die. There they accumulate as **foraminiferal oozes,** sandy and silty sediments composed of foraminiferal shells (Figure 17.35). Other carbonate oozes are made up of shells of different microorganisms, called *coccoliths.*

Foraminiferal and other carbonate oozes are abundant at depths of less than about 4 km, but they are rare on the deeper parts of the ocean floor. This cannot be because of a lack of shells, for the surface waters are full of them everywhere and the living foraminifera are unaffected by the bottom far below. The explanation for the absence of carbonate oozes below a certain depth, called the **carbonate compensation depth (CCD),** is that the shells dissolve in deep seawater (Figure 17.36). Because of the nature of the physical circulation of the oceans, the deeper waters of the ocean differ from shallower waters in three ways:

- They are colder—colder, denser polar waters sink beneath warmer tropical waters and travel toward the equator along the bottom.
- They contain more carbon dioxide—not only do colder waters absorb more carbon dioxide than warmer waters, but any organic matter they are carrying tends to be oxidized to carbon dioxide during their long circulation travels.
- They are under higher pressure—this pressure results from the greater weight of overlying water.

These three factors make calcium carbonate more soluble in deep waters than in shallow ones. As the shells of dead foraminifera fall to the bottom below

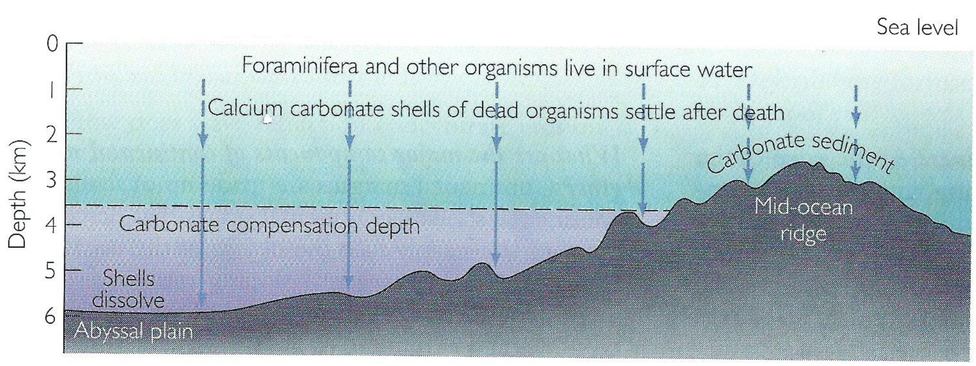


FIGURE 17.36 The carbonate compensation depth is the level in an ocean below which the calcium carbonate of foraminifera and other carbonate-shelled organisms that have settled from surface waters dissolves. As the shells of dead organisms settle into the deep waters, they enter an environment undersaturated in calcium carbonate and so dissolve.

the CCD, they enter an environment undersaturated in calcium carbonate, and they dissolve.

Another kind of biochemically precipitated sediment, **silica ooze,** is produced by sedimentation of the silica shells of diatoms and radiolaria. Diatoms are green, unicellular algae found in abundance in the surface waters of the oceans. Radiolaria are unicellular organisms that secrete shells of silica instead of calcium carbonate. After burial on the seafloor, silica oozes are cemented into the siliceous rock, chert.

Some components of pelagic sediments are formed by chemical reactions of seawater with sediment on the seafloor. The most prominent examples are manganese nodules, which are black, lumpy accumulations ranging from a few millimeters to many centimeters across. These nodules cover large areas of the deep ocean floor, as much as 20 to 50 percent of the Pacific. Rich in nickel and other metals, they are a potential commercial resource if we find some economical way to mine them from the seafloor.

Differences in the Geology of Oceans and Continents

Our studies of the ocean floor have given us an appreciation of the differences between continental geology and submarine geology. Whereas tectonics and erosion shape the continents, volcanism and sedimentation predominate in the ocean. Volcanism creates island groups, such as the Hawaiian Islands, in the middle of the oceans; arcs of volcanic islands near deep oceanic trenches; and mid-ocean ridges. Sedimentation shapes much of the rest of the ocean floor. Soft sediments of mud and calcium carbonate blanket the low hills and plains of the sea bottom and accumulate on oceanic plates as they spread from mid-ocean ridges. As the plates move farther and farther from a ridge, they accumulate more and more sediment. Eventually the plates are swallowed by subduction zones, which destroy the oceanic sediment record by metamorphism and melting.

The oceans have no folded and faulted mountains like those on the continents. Instead, plate-tectonic deformation is restricted to the faulting and volcanism found at mid-ocean ridges and at subduction zones. Weathering and erosion are much less important in the oceans than on the land because there are no efficient fragmentation processes, such as freezing and thawing, or major erosive agents, such as streams. Deep-sea currents can erode and transport sediment but cannot effectively attack basaltic plateaus or hills.

Because tectonic deformation, weathering, and erosion are minimal over much of the seafloor, many more details of the geologic record are preserved in layers of oceanic sediment than in continental sediments. But the oldest parts of the rock record are continuously erased by subduction. The oldest sediments preserved on today's ocean floor are Jurassic, about 150 million years old; they lie at the western edge of the Pacific Plate. In the next million years, they too will disappear down a subduction zone. It takes about 150 million years, on average, for the crust created at mid-ocean ridges to spread across an ocean and come to a subduction zone.

This brief survey of the oceans shows that they are geologically complex, with distinctive structures, topographies, and sediments. Our knowledge of the oceans is still in its infancy; much of what we know was discovered only in the past few decades. Much more remains to explore as we invent new ways to sound, map, and sample the ocean floor.

Summary

What processes shape shorelines? At the edge of the sea, waves and tides, interacting with tectonics, control the formation and dynamics of shorelines, from beaches and tidal flats to uplifted rocky coasts. Waves are generated by winds blowing over the sea; as the waves approach the shore, they are transformed into breakers in the surf zone. Wave refraction results in longshore currents and longshore drift, which transport sand along beaches. Tides, generated by the gravitational attraction of the Moon and Sun on the water of the oceans, are agents of sedimentation on tidal flats.

What are the major components of continental margins? Continental margins are made up of shallow continental shelves; continental slopes that descend more or less steeply into the depths of the ocean; and continental rises, gently sloping aprons of sediment deposited at the lower edges of the continental slopes and extending to abyssal plains farther out in the ocean. Waves and tides affect the continental shelves, but continental slopes are shaped by turbidity currents, deep-water currents formed as slumps and slides on the continental slope create turbid suspensions of muddy sediment in bottom waters.

Turbidity currents also produce submarine fans, submarine canyons, and abyssal plains. Active continental margins form where oceanic lithosphere is subducted beneath a continent, and passive continental margins form where rifting and seafloor spreading carry continental margins away from plate boundaries.

How is the deep seafloor formed? The deep seafloor is constructed by volcanism at mid-ocean ridges and at oceanic hot spots such as Hawaii and by deposition of fine-grained clastic and biochemically precipitated sediments. Mid-ocean ridges are the sites of seafloor spreading and the extrusion of basalt, which produces new oceanic lithosphere. Deep-sea trenches form as oceanic lithosphere is pulled downward into a subduction zone. Isolated, submerged seamounts and guyots, volcanic islands, plateaus, and abyssal hills are all accumulations of volcanic rock, most of which are mantled by sediment. Coral reefs form as volcanic islands are fringed with reefs constructed by organisms, which then continue to build the island and keep it at the surface when the volcano becomes extinct and subsides below sea level. Pelagic sediments consist of reddish-brown clays and foraminiferal and silica oozes composed of the biochemically precipitated calcium carbonate and silica shells of microscopic organisms living in surface ocean waters.

Key Terms and Concepts

world ocean (p. 420)
coast (p. 420)
shoreline (p. 421)
swell (p. 421)
wavelength (p. 422)
wave height (p. 422)
period (p. 422)
surf (p. 422)
surf zone (p. 422)
swash (p. 423)
backwash (p. 423)
wave refraction (p. 424)
longshore drift (p. 424)
longshore current (p. 424)
tide (p. 426)
spring tide (p. 427)
neap tide (p. 427)
tidal surge (p. 427)
flood tide (p. 428)
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tidal flat (p. 428)
offshore (p. 428)
foreshore (p. 428)
backshore (p. 429)
stack (p. 432)
wave-cut terrace (p. 432)
spit (p. 433)
barrier island (p. 433)
estuary (p. 435)
continental shelf (p. 437)
continental slope (p. 437)
submarine canyon (p. 437)
continental rise (p. 437)
abyssal plain (p. 438)
seamount (p. 438)
continental margin (p. 440)
passive margin (p. 440)
active margin (p. 440)
turbidity current (p. 441)
submarine fan (p. 444)
turbidite (p. 444)
guyot (p. 447)
atoll (p. 448)
fringing reef (p. 448)
pelagic sediments (p. 449)
foraminiferal ooze (p. 451)
carbonate compensation depth (CCD) (p. 451)
silica ooze (p. 452)

Exercises

1. How are ocean waves formed?
2. How does wave refraction concentrate erosion at headlands?
3. Where do turbidity currents form?
4. How has human interference affected some beaches?
5. Where and how is the deep seafloor created by volcanism?
6. What plate-tectonic process is responsible for deep-sea trenches?
7. Describe the formation of a coral reef in the open ocean.
8. What are pelagic sediments?
9. Along what kinds of continental margins do we find broad continental shelves?
10. What features of the seafloor are associated with plate divergences and with plate convergences?
11. What processes in the ocean are responsible for foraminiferal and silica oozes?