Chapter 5

Sedimentary Environments

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The stratigraphic record reveals remarkably detailed pictures of the settings where sediments accumulated long ago, enabling geologists to reconstruct ancient environments and, more broadly, paleogeography—the geography of the past. The goal is not only to learn about the distribution of land and sea at a particular time, but also to identify localized environmental features of the past, such as deserts, lakes, river valleys, lagoons, and submarine shelf breaks. In most instances, geologists can learn not only where a valley was located, for example, but also what kind of river occupied it—perhaps one consisting of many small, intertwining channels choked with bars of gravel and sand, or one flowing along a single broad, winding channel. Frequently geologists can also “read” from the sedimentary record whether the terrain that once bordered an ancient river was a dry, sparsely vegetated plain or a swamp densely populated by water-loving trees and undergrowth. In this and other ways, geologists reconstruct ancient climates and habitats.

The identification of ancient sedimentary environments also provides geologists with a framework within which to interpret life of the past. Although we can learn some aspects of how an organism lived by studying its fossil remains alone, a fuller understanding of that species can come only when its habitat is taken into consideration. It was once widely believed, for example, that the largest dinosaurs were too big to be fully terrestrial and therefore must have spent much of their time in water, like hippopotamuses. The stratigraphic record contradicts this idea: the fossilized bones of these giant creatures are frequently found in sedimentary rocks that represent nonaquatic environments.

Furthermore, by understanding the environmental relationships of sedimentary rocks, geologists can often predict where sedimentary deposits may harbor such natural resources as petroleum and natural gas, or where economically valuable sedimentary bodies, such as deposits of halite and gypsum, may exist. Petroleum and natural gas tend to accumulate in porous sediments, such as clean sands deposited along ancient shorelines or rivers, and in ancient limestone reefs constructed in shallow seas by corals or other organisms.

Geologists rely heavily on actualism to reconstruct the environments in which sedimentary rocks accumulated. In other words, they study patterns of deposition in modern settings in order to recognize ancient deposits that formed in similar environments. Because sediments deposited in modern settings tend to be buried quickly beneath younger strata, geologists often have to excavate, tunnel, or core before they can observe those sediments. Coring—whereby a tube is driven into a sedimentary deposit and then withdrawn—is especially useful for examining deposits at the center of a lake, lagoon, or deep ocean. Cores of sediment thus extracted can indicate the sequence of deposition at the site and can provide a three-dimensional picture of the deposit. Similarly, geologists who wish to examine a meandering river's depositional record must either dig one or more pits in the valley floor adjacent to the channel or sample the floor by means of coring.

Although some sedimentary features provide highly reliable information about the nature of the environments in which they formed, many others offer ambiguous testimony. Most coal deposits, for example, represent swamps choked with vegetation—but such swamps typically border either the banks of rivers or the shores of marine lagoons. To determine which environment pertains, geologists must evaluate the beds that lie above or below coal deposits. If sediments containing fossils of marine animals are found, it is likely that a marine lagoon, not a river, lay adjacent to the swamp or marsh in which the coal formed. Such vertical stacking of sedimentary features often provides the key to recognizing ancient environments.

Nonmarine Environments

We will begin our exploration of sedimentary clues to ancient environments by considering depositional settings on the surfaces of continents. Then our focus will shift to the margins of continents, and finally to the marine realm.

Ancient soils can point to past climatic conditions

Soil can be defined as loose sediment that contains organic matter and has accumulated in contact with the atmosphere rather than under water. Soil rests either on sediment of some sort—such as sand or mud—or on rock. It supplies essential nutrients and provides physical support for plants, and it serves as a habitat for many organisms. Soils develop largely through weathering processes and the decay of plant material. Topsoil, the upper zone of many soils, consists primarily of sand and clay mixed with humus. Humus, the organic matter that gives topsoil its dark color, is derived from the decay of plant debris by bacteria, and it supplies nutrients for other plants in turn; thus it has an important position in the cycling of materials through terrestrial ecosystems.

Soils form in a variety of environments throughout the world—in tropical rain forests, in arid regions, and even on mountaintops—and they can often be found buried, albeit chemically altered, within thick sequences of ancient sediment. The type of soil that forms depends in part on climatic conditions. For example, in warm climates that are dry for part of the year, evaporation of groundwater causes calcium carbonate to precipitate as nodular or massive deposits known as caliche (Figure 5-1). In moist tropical climates, warm waters percolate through the soil, destroying humus by oxidizing its components (combining them with oxygen). Silicate minerals in such areas also break down quickly, producing oxisols: soils rich in iron oxides, which give them a rusty red color, as well as aluminum oxides. Most soils are destroyed by erosion. In addition, ancient buried soils can be exceedingly difficult to recognize and to interpret, partly because their original chemical components are often altered beyond recognition. One place where an ancient soil may be found,
however, is beneath an unconformity. Ancient soils are most likely to be found beneath unconformities that have been recognized on the basis of other criteria. Plant roots also provide clues to the presence of ancient soils. Similarly, burrows made by animals such as insects and rodents are diagnostic features. Certainly the most unusual of these excavations are the structures known as “devil’s corkscrews,” which are actually burrows that beavers of an extinct species dug with their teeth in the Oligocene and Miocene soils of Nebraska (Figure 5-2). Skeletons of these beavers have been found in the burrows, and scratches on the burrow walls match their front teeth! The fact that these animals lived as far as 10 meters (33 feet) below ground level indicates that the level of standing water in the ancient soil stood below this depth; had it been higher, the beavers would have drowned.

**Freshwater lakes and glaciers leave clues to ancient climates**

Freshwater lakes and glaciers serve as reservoirs of water on land. Both leave distinctive features in the geologic record that provide indications of ancient climatic conditions.
Freshwater lake environments At no time in geologic history have lakes occupied a large fraction of Earth’s surface. However, because lakes form in basins that lie at lower elevations than most soils, lake deposits are much more likely than soils to survive erosion.

Evidence of large freshwater lakes in the geologic past indicates abundant precipitation in a region because such lakes must be fed by substantial runoff from the land. In addition, because of the high heat capacity of water, large lakes tend to stabilize the temperature of nearby land areas, reducing annual fluctuations.

The sediments around the margins of a freshwater lake tend to be coarser than those at its center, partly because the current of a stream or river slows where it meets the waters of a lake, dropping its load of coarse sediment near the shore. Furthermore, the wind-driven waves on a deep lake’s surface touch bottom only where they approach the shore, winnowing the sediment there and driving clay-sized particles into suspension. These particles later settle to the bottom toward the lake’s center. Some lakes receive almost all of their coarse sediment during a moist season, and only fine-grained sediment accumulates during the rest of the year. The resulting pairs of coarse-grained and fine-grained layers are known as varves.

Fossils are valuable tools for distinguishing lake sediments from marine sediments. Although fish fossils are found in both kinds of deposits, the presence of exclusively marine fossils, such as corals or echinoderms, provides strong evidence that ancient sediments did not originate in a lake. Furthermore, because burrowing animals are not as abundant in lakes as they are in many marine environments (and because waves and currents in lakes are generally weak), the fine-grained sediments that accumulate in the centers of lakes are likely to remain well layered—just as well layered as when they were laid down (Figure 5-3).

Another clue to the identification of lake deposits is close association with other nonmarine deposits, such as river sediments. It would be highly unusual to find lake deposits directly above or below deep-sea deposits, or even above continental-shelf sediments, unless the two types of deposits were separated by an unconformity.

Glacial environments Sedimentary features associated with continental glaciers are excellent indicators of cold climates. Glaciers that form in mountain valleys seldom leave enduring geologic records because the mountains stand above the surrounding terrain and therefore tend to be eroded rapidly. Continental glaciers, however, leave legible records that survive for hundreds of millions of years. The geologic record reveals that ice sheets have spread over broad geographic areas several times during Earth’s history. In fact, we are living during an interval of continental glaciation. Modern examples of continental glaciers include the one now occupying most of Greenland (see Figure 4-13) and the even larger ones covering nearly all of Antarctica.

As glaciers move, they erode rock and sediment, transporting both in the direction of flow. Rocks embedded in the base of a glacier commonly leave deep scratches in the underlying rock, which serve to record the direction of the glacier’s movement. When such scratches are found in an area now free of glaciers, it is safe to assume that a glacier passed that way in the distant past (Figure 5-4).

Glaciers leave records not only of erosion but also of deposition. Some of the mixture of boulders, pebbles, sand, and mud scraped up by a moving glacier is deposited as the glacier advances, and some is deposited when it melts. This unsorted, heterogeneous material is called till (Figure 5-5); when lithified, it is known as tillite. At the farthest reach of a glacier’s advance, till plowed up in front of the glacier is left standing in ridges known as moraines. In front of a moraine, sediment from a melting glacier is often deposited by streams of meltwater issuing from the front of the retreating mass of ice. Here the sediment tends to be sorted by size into layers of gravel, cross-bedded sand, and mud, forming well-stratified glacial material known as outwash.
the winter layer darker than the summer layer. Each pair of layers of varved sediment thus represents a single year’s deposition, so geologists can count the number of years represented by a series of layers. In some areas, thousands of years of deposition have been tallied in this manner.

When a glacier encroaches on a lake or ocean, pieces of it break loose and float away as icebergs (see Figure 4-14). As these chunks of glacial ice melt, the sediment they carried sinks to the bottom of the lake or ocean, creating a highly unusual deposit in which pebbles or even boulders rest in a matrix of finer sediments. Unlike the tightly packed, coarse material that characterizes glacial till, these dropstones occur either singly or scattered throughout the matrix. Very few natural mechanisms other than this ice rafting bring large stones to the middle of a lake or to a seafloor far from land.

### Deserts and arid basins accumulate salt and sand

Like ancient glacial environments, dry environments of the distant past can be identified from diagnostic traits of sedimentary rocks. Desert soils contain little organic matter because dry conditions support little vegetation, the source of the organic matter in soils. The rain that occasionally falls in deserts leads to erosion and deposition of sediment, and temporary streams carry the chemical products of weathering to desert basins. The subsequent precipitation of evaporite minerals in these basins sets arid regions apart from those with moist climates. In moist regions, permanent streams flow great distances nearly 2 billion years older (B). The sediments in B also contain a pebble of metamorphic rock, known as a dropstone, that apparently dropped from floating ice. (A. Courtesy Tom Hooyer, University of Wisconsin-Milwaukee; B. Courtesy Tim Howe.)
without being absorbed into the soil or evaporating into the air, so they are frequently capable of reaching the ocean. As a consequence, we speak of most humid regions as having **exterior drainage**, meaning that water and sediment are transported beyond their borders. Runoff from arid regions, in contrast, is too sparse and intermittent to form permanent streams and rivers, and often the result is **interior drainage**—a pattern in which streams die out through evaporation, seepage into the dry terrain, or drainage into lakes. Lakes in areas with interior drainage, known as **playa lakes** (Figure 5-7), also tend to be temporary, and when they shrink by evaporation, evaporites precipitate from them.

In dry, sparsely vegetated regions, the wind may pile loose sand into hills of sand called **dunes**. Dunes occupy less than 1 percent of some deserts but sometimes form magnificent landscapes (Figure 5-8). Where the wind blows across loose sand, a dune can begin to form over any obstacle that creates a wind shadow in which sand can accumulate. Figure 5-9A shows how a dune tends to "crawl" downwind as sand from the upwind side moves over the top and accumulates on the downwind side. As the prevailing wind direction shifts back and forth, the direction in which the dune migrates shifts as well. A shift in wind direction usually leads to the truncation of preexisting deposits, which often causes a new set of beds to accumulate on a curved surface cut through older sets. The result is called **trench cross-stratification**. Figure 5-9B shows an idealized cross section through a dune, and Figure 5-9C shows a real section through an enormous lithified dune more than 200 million years old.

Dunes are familiar sights not only in desert terrain, but also landward of the sandy beaches that border oceans and large lakes. For this reason, a geologist must consider other indicators to determine the setting in which lithified sand dunes formed. However, even after a geologist identifies an arid region as the setting for ancient dunes, there is work to be done. Dry climates are widespread in the trade wind belt, but are also found in the rain shadows of mountains and in inland regions, far from oceans. To determine which of these settings was the site of an arid environment recorded in ancient rocks, geologists...
The internal structure of a sand dune. A. The windstream becomes compressed just above a dune and consequently increases in velocity. The dune ceases to grow taller when its height becomes so great that it causes the windstream to move rapidly enough to transport sand. As sand passing over the dune accumulates on the steep leeward slope, the dune begins to “crawl” in that direction. B. This cross section of a dune shows the cross-bedding that results from shifts in the wind. As the wind direction changes, the shape of the dune is altered by removal of sand, and a new leeward slope forms. C. Cross-stratified dune deposits of the Jurassic Navajo Sandstone in Arizona. (C. Crisma/Yetta/Getty Images.)

Many of the features typical of an arid basin can be seen in Death Valley, California (see Figures 5-7, 5-8, 5-10, and 5-11). Temporary streams carry sediment down valleys incised into the naked rocks of nearby highlands to form low cone-shaped structures called alluvial fans, which spread out onto the floor of the valley (Figure 5-10). These structures form where a mountain slope meets the valley floor, causing the streams to slow down and drop much of their sediment. Alluvial fans consist of poorly sorted sedimentary particles that range from boulders to sand near the source area and from sand to mud on the lower, gentler slopes. Much of this material is deposited by debris flows, which are downward movements of loose material under the influence of gravity. Most alluvial fans include broad deposits of coarse, cross-stratified sediments laid down by a complex network of channels, or braided streams, that carry water and sediment during the infrequent rainy intervals. After a heavy rain or snowmelt in the mountains, water gushing onto a fan from a steep valley carries so much sediment that sand and gravel clog portions of the initial channel, dividing it into numerous smaller channels that connect and divide in complex patterns.

The braided streams formed in this way lead to the center of the basin, where their occasional flow may form a temporary playa lake (see Figure 5-7). As the lake waters dry up, evaporite minerals accumulate. Similar minerals also accumulate on those parts of the basin floor where groundwater seeps to the surface and evaporates. The evaporites of Death Valley are composed primarily of halite, gypsum, and anhydrite. Alternate wetting and drying in this basin produce large polygonal mudcracks in many areas (Figure 5-11). Around the margins of these “salt pans,” calcium carbonate deposition forms caliche.
Large mudcracks in Death Valley, California. These mudcracks have formed quite recently in soft sediment, and many are about 1 meter (3 feet) across. (Peter L. Kresan.)

Braided and meandering rivers deposit sediment in moist regions

In areas that receive abundant rainfall, the precipitation normally creates exterior drainage: small streams meet to form larger streams, which in turn flow into still larger rivers. Many small rivers terminate at lakes, but most large rivers reach the sea. Figure 5-12 summarizes the sequence of aquatic environments through which water typically passes as it moves from the headwaters of river systems in hills or mountains to the deep sea, transporting and depositing sediment in the process. In the remainder of this chapter we will investigate the diagnostic features of ancient deposits formed in each of these environments—features that enable us to recognize each type of deposit as far back in the geologic record as a few hundred million or even 2 billion or 3 billion years.

Alluvial fans and braided-stream deposits Sediment in the form of alluvial fans accumulates at the feet of mountains and steep hills in moist regions, just as it does in arid regions. In moist climates, however, water often spreads these coarse sediments farther from their source, producing fans that slope more gently than those of dry basins. Alluvial fans in moist regions nonetheless slope steeply enough to remain poorly vegetated, and the large volumes of coarse sediment that arrive at times of heavy rainfall or snowmelt produce braided streams. Braided streams also form in front of mountain glaciers that experience substantial melting in summer, as the meltwaters are choked with sand and gravel plowed up by the flowing ice (Figure 5-13).

Meandering rivers In gently sloping terrain far from uplands, rivers generally occupy solitary channels that wind back and forth like ribbons (Figure 5-14). Unlike braided streams, these meandering rivers—such as the Mississippi or the Thames—are not choked with sediment because sediment is supplied to them slowly in relation to the rate at which the water flows. Any irregularity in the local terrain causes the river’s path to curve. Because the water flows least rapidly near the inside of a bend and most rapidly near the outside, the river tends to cut into the outer
A braided stream. Bars of gravelly sediment divide the flow into numerous winding channels. Most of the water and sediment in this stream emerge from the base of melting mountain glaciers, which are located upstream. (David Wall/Alamy)

Deposition of sediment by a meandering river. A. At each bend, the river migrates outward. The current flows most rapidly on the outside of a bend (long arrow), so the water cuts into the outer bank. On the inside of the bend, where the current moves more slowly, sand accumulates to form a point bar. As the river channel migrates outward, the point-bar sands advance over the coarser, cross-bedded sands deposited within the original channel. Pebbles often accumulate at the base of the channel. Muddy floodplain deposits, which form when the river floods its banks, migrate in their turn over the point-bar sands. As a result of this shifting of depositional environments, coarse sediments at the base of the sequence grade upward into fine sediments at the top. B. Cyclical meandering-river deposits of Oligocene age in the Sespe Formation, California. In each cycle, coarse-grained sandstone (the light-colored, thickly bedded deposits), which formed in the river channel, grades upward into fine-grained sediments (red), which accumulated on the floodplain. The thickest layer is about 8 meters (26 feet). (B from Douglas J. Cant, in P. A. Scholle and D. Spearing, eds., *Sandstone Depositional Environments*, AAPG Memoir 31, American Association of Petroleum Geologists, Tulsa, OK, 1982; B: Douglas J. Cant.)
Point-bar deposits usually consist of sand. Most of this sand is cross-bedded because large ripples migrate along the riverbed in the course of time. In deeper water, where the current is stronger, the sediment in the river channel is coarser (Figure 5-15B); pebbles are often found along with coarse sand in the deepest part of the channel.

Because mud consists of very fine particles, it tends to move downstream without settling within a meandering river’s channel. When the river overflows its banks, however, it carries fine sediment laterally to the adjacent lowlands, known as floodplains, which may include moist vegetated areas known as backswamps. Here, the spreading floodwaters flow slowly, allowing mud to settle before they recede. In keeping with the normal pattern of sediment deposition, these floodwaters move progressively more slowly as they flow away from the channel, so they tend to drop the coarser portion of their suspended sediment before they spread far from the channel. Sand is therefore dropped first, followed by silt, and together these deposits form a gentle ridge, or natural levee, alongside the channel. Because natural levees and floodplains are inundated only periodically, their surfaces tend to dry out and crack. Many mudcracks formed in this way have been preserved in the stratigraphic record. Levees and floodplains also become populated by moisture-loving plants, which may leave traces of their roots in the rock record or even form deposits that eventually turn into coal.

The vertical sequence in which sediments are deposited by a meandering river is shown in Figure 5-15—from coarse channel deposits at the base and cross-beded point-bar sands in the middle to muddy floodplain deposits at the top. Levee sediments sometimes lie between the point-bar and floodplain deposits. This coarse-to-fine sequence can be thought of as forming a composite depositional unit. It forms because, as the channel migrates laterally, the point bar builds out over deeper gravels, and the floodplain shifts over older point-bar deposits. The meandering-river sequence shown in Figure 5-15 illustrates Walther’s law, which states that when depositional environments migrate laterally, sediments of one environment come to lie on top of sediments of an adjacent environment.

In a broad basin that happens to be subsiding (sinking relative to surrounding terrain), a river may migrate back and forth over a large area many times, piling one coarse-to-fine composite depositional unit on top of another as it goes (Figure 5-15B). Each of these composite units, or sedimentary cycles, lies unconformably on the one beneath it because the channel in which the basal deposits accumulate removes the uppermost sediments of the preceding cycle as it migrates. Sometimes a channel cuts deeply, removing not only the uppermost deposits of the preceding cycle, but some lower ones as well. Thus many preserved cycles of migrating channels are really only partial cycles.

Two final points deserve mention. First, not every river can be assigned to the braided or meandering category. Some rivers have segments or branches that are braided and others that are meandering. Second, no river in a moist region deposits its entire load of sediment in its valley. A river ultimately discharges its water and remaining sediment into a lake or the sea (see Figure 5-12).

Marginal Marine and Open-Shelf Environments

Along the edge of the sea, sediment accumulates under the influence of both fresh water and seawater. Thus freshwater plants may form peat deposits in a marsh near the coast, while waves that break nearby deposit sand and bury shells of marine animals along a beach exposed to the open ocean. Continental-shelf environments are typically dominated either by carbonate sediments or by siliciclastic sediments.

A delta forms where a river meets the sea

Where a river empties into either a lake or the sea, its current dissipates, and it often drops its load of sediment in a fanlike pattern. The depositional body of sand, silt, and clay that is formed in this way is called a delta because of its resemblance to the Greek capital letter Δ. Most of the large deltas that have been well preserved in the geologic record formed in areas where sizable rivers emptied into ancient seas.

Like all moving water, as river water mixes with standing water and begins to slow down, it loses sand first. Silt, being finer, spreads farther from the mouth of the river, and clay is carried even farther. The typical result is a delta structure that includes delta-plain, delta-front, and prodelta deposits (Figure 5-16). Delta-plain beds, which consist largely of sand and silt, are nearly horizontal except where they are locally cross-beded. Some delta-plain deposits accumulate within river channels. As the river slows on the surface of the delta, sand builds up on the bottom, causing the channel to branch repeatedly into smaller channels that radiate out from the mainland. These distributary channels are floored by cross-beded sands. Sand also spills out from the mouths of the channels, forming shoals and sheetlike sand bodies along the delta front. Between the distributaries and separated from them by natural levees are swamps, which are sometimes dotted by lakes. Here, as in the backswamps of meandering rivers, muds accumulate and marsh plants often grow, contributing to future coal deposits.
Delta-front beds slope seaward from the delta plain, usually lying in waters too deep to be agitated by wind-driven surface waves. These beds consist largely of silt and clay, which can settle under these quiet conditions. Because they lie fully within the marine system, delta-front muds harbor marine faunas that often leave fossil records, but these muds usually contain fragments of waterlogged wood as well. In fact, the presence of abundant fossil wood in ancient subtidal marine muds testifies to the presence of both land and a river system near the site, and such muds usually represent deltaic deposits.

Spreading seaward at a low angle from the lowermost delta-front deposits are the prodelta beds, which consist of clay. Even during floods, the fresh water that spreads from distributary channels slows down so abruptly that it loses its silt on the delta front. It is partly because fresh water is less dense than seawater that clay is carried far from the mouths of distributary channels. Because the fresh water floats on top of the denser seawater, it does not mix in quickly; instead, it spreads seaward for some distance, still carrying much of its clay.

As a delta progrades, or grows seaward, the relatively coarse deposits of the delta plain build out over the finer-grained delta-front beds, in accordance with Walther’s law (see Figure 5-16), and the delta-front beds build out over the still finer-grained prodelta beds. The result is a sequence of deposits that coarsens toward the top.

The famous Mississippi River delta spills into the Gulf of Mexico in an area that is protected from strong wave action. As a result, the delta projects far out into the sea. Because construction of the delta from river-borne sediment has prevailed decisively over the destructive forces of the sea, this type of delta is sometimes called a river-dominated delta. The growing portion, or active lobe, of such a delta is the functioning distributary channels (Figure 5-17). Many previously active lobes can be identified in the delta-plain portion of the Mississippi delta, and these lobes, which have been dated by the carbon 14 method (described in Chapter 6), provide a history of deltaic development during the past several thousand years (Figure 5-18). Depositional activity (or lobe growth) periodically shifted when floods caused the river to cut a new channel and to abandon the previously active channel and its distributaries.
FIGURE 5-18  Holocene lobes of the Mississippi River delta in the southern United States. The positions of seven major depositional lobes record the shifting positions of sediment discharge into the Gulf of Mexico. All of these lobes were deposited over the last 7000 years, after sea level came close to its present position following the last ice age. Lobe 7 is the currently active lobe. The oldest lobe (1, where Atchafalaya Bay is located) has subsided over time to the degree that it is now mostly submerged.

The fate of an abandoned deltaic lobe is the key to the stratigraphic sequence of a river-dominated delta. An abandoned lobe gradually sinks, for two reasons: first, the sediments of which it is formed compact under their own weight, and second, the lobe is part of the entire deltaic structure, which is constantly sinking as a result of the isostatic response of the underlying crust to the weight of the continually growing mass of sediment. A younger lobe eventually grows on top of an abandoned lobe, with each consisting of the typical upward-coarsening sequence. The result is an accumulation of sedimentary cycles that differ markedly from those of meandering rivers, which, as you will recall, become finer-grained toward the top.

Some deltaic cycles in the rock record lack tops because sediment was eroded away before another cycle was superimposed (Figure 5-19). Because the building of a river-dominated delta is limited to the active lobe at any given time, and because older exposed beds erode easily, a delta can seldom be traced very far laterally in the rock record. When preserved, the porous sand bodies in the upper parts of deltaic cycles may serve as reservoirs for petroleum or natural gas.

Changes in the rate at which a delta sinks or is supplied with sediment can alter the size of the active lobe. Today such changes are causing the active lobe of the Mississippi delta to shrink rapidly, with alarming consequences. The building of artificial levees along the Mississippi River and of dams along its tributaries have reduced the rate of deposition of sediment on the delta. In addition, the removal

FIGURE 5-19  Five deltaic cycles. Cycle 3 is a complete cycle: within it, prodelta clays grade upward to delta-plain sands. Each cycle represents an accumulation of sediments resulting from the seaward growth of a deltaic lobe. The thickness of a cycle reflects the size of the delta. Many deltaic cycles are 5–10 meters (16–33 feet) thick.
of groundwater for human use in the region of the delta has increased the rate of subsidence of the delta fivefold. Because of these two factors, the Louisiana coast is now losing about 100 square kilometers (40 square miles) of land every year. Furthermore, as the marine waters of the Gulf of Mexico encroach farther and farther inland over the delta, they are drowning the swamps, which provide food and shelter for nearly 30 percent of the annual seafood harvest of the United States and support the nation’s largest population of waterfowl.

**Lagoons lie behind barrier islands of sand**

Deltas of large rivers occupy only a small percentage of the total shoreline of the world's oceans. Where they are absent, some long stretches of shoreline are fringed by barrier islands, composed largely of clean sand piled up by waves. Although some barrier islands extend laterally from deltas, most derive their sand not from neighboring rivers, but from the marine realm. They are built up as waves and the shallow currents that flow along the coast, called longshore currents, winnow sediments and sweep sand parallel to the shoreline. Where the beach of a barrier island is washed by breaking waves, deposits tend to have nearly horizontal bedding, but often dip gently seaward. Cross-bedding develops in areas where the beach surface is gently irregular and changes from time to time. Windblown sand often accumulates behind the beach as sand dunes, but in time these dunes are often eroded away.

**Lagoons lie behind long barrier islands, such as those that border the Texas coast (Figure 5-20).** Protected from strong waves, lagoons trap fine-grained sediment and are usually floored by muds and muddy sands. Small rivers often build deltas along the landward margins of lagoons. A barrier island and the lagoon behind it form a barrier island–lagoon complex.

Barrier islands often form chains with tidal channels separating adjacent islands. Tidal currents pass through these channels and deposit cross-bedded tidal deltas within the lagoon. Other depositional environments are also found along the shores of lagoons. Among them are tidal flats—formed of sand or muddy sand—whose surfaces are alternately exposed and flooded as the tide ebbs and flows. High in the intertidal zone, above the barren tidal flats, marshes fringe one or both margins of many lagoons. Here plant debris accumulates rapidly and decomposes to form peat or, after long burial, coal.

Fresh water from rivers and streams tends to remain trapped in coastal lagoons for some time, so the waters of lagoons in moist climates are often brackish. The salinity of these waters at any given time depends on the rate of freshwater runoff from the land, which varies during the course of the year. Laguna Madre of Texas is typical of lagoons found in warm, arid climates (see Figure 5-20). The ponded waters of this long lagoon are hypersaline because they receive little fresh water from rivers and experience a high rate of evaporation. Whether lagoons are brackish or hypersaline, their abnormal and fluctuating salinity excludes many forms of marine life. As a result, the fossil faunas in the ancient sediments of lagoons are not very diverse; those species that are present, however, often occur in large numbers. Usually among them are burrowers, such as segmented worms, that disturb the muddy lagoonal sediments, leaving them mottled and largely devoid of bedding structures (see Figure 2-27).

When a barrier island–lagoon complex receives sediment at a sufficiently high rate, it progrades—that is, it migrates seaward—like the active lobe of a delta. Unlike the migration of a delta, however, this progradation takes place along a broad belt of shoreline (Figure 5-21).

![Figure 5-20](image)
**Figure 5-20** Barrier islands along the Texas coast. Tides here are weak, and there are few tidal channels or passes, so large lagoons lie behind these Texas barrier islands.

![Figure 5-21](image)
**Figure 5-21** The stratigraphic sequence produced when a barrier island–lagoon complex progrades. Sediments of the lagoon and of the adjacent marshes and tidal flats are superimposed on the beach sands of the barrier island.
As the shoreline migrates seaward, marsh and tidal-flat deposits prograde over the sediments of the lagoon and its associated tidal channels. All these sediments, in turn, build out over the sand of the barrier islands and over the tidal deltas and marshes behind them. Thus the horizontal sequence of depositional environments (barrier island, marsh or tidal delta, lagoon, tidal flat, and marsh) comes to be represented by a corresponding vertical sequence of sedimentary deposits, in accordance with Walther’s law.

**Open-shelf deposits include tempestites**

Seaward of barrier islands, continental shelves display a variety of physical conditions and therefore produce a variety of sedimentary deposits. On open shelves where tides produce strong currents and sand is abundant, the currents may pile the sand into large ridges or dunelike structures. On shelves where waves have a stronger effect than tidal currents, wave motion tends to flatten the bottom, and the sand spreads out in sheets. Many storm-produced shelf deposits preserve low-angle “hummocky” cross-stratification produced in deeper water during large storms.

On quieter shelves, mud or muddy sand accumulates most of the time, but storms occasionally produce tempestites, which are sandy beds that are usually a few centimeters thick. A storm that powders a coastline may produce waves that pile up water carrying sand that they have scoured from the shallow seafloor. After the storm passes inland, the piled-up water flows seaward, and as it loses velocity, it deposits the suspended sediment on the shelf in the form of a tempestite. The occasional deposition of tempestites on a normally quiet shelf produces a succession of sandy beds in a body of finer-grained sediment (Figure 5-22). Individual tempestites are commonly graded, having formed as sand accumulated before silt or mud, and the basal sand is usually cross-bedded.

Animal burrows are abundant in most shelf sediments, and as we will soon see, particular skeletal fossils also point to an offshore, open marine environment.

**Fossils serve as indicators of marine environments**

Ancient sediments deposited within and seaward of barrier island–lagoon complexes often yield fossils that help geologists to recognize particular depositional environments. Figure 5-23 depicts an example in Wales, west of central England. Here, in rocks of mid-Silurian age, fossil communities of marine invertebrates are arrayed roughly parallel to the ancient shoreline. Adjacent to that shoreline is a narrow zone of fine-grained sediments, in which the inarticulate brachiopod *Lingula* is especially abundant, but only a few other species are present (Figure 5-23A). Presumably the water here was brackish: *Lingula*, a living fossil genus (see Figure 3-28B), today tolerates nearshore environments of brackish and variable salinity where few other species are able to live.

Seaward of the zone where *Lingula* predominates in the Silurian deposits is a more diverse fossil community, adapted to the more stable conditions of the center and seaward margin of a lagoon (Figure 5-23B). Sandy deposits representing a barrier island have not been preserved, but we can infer that one was present because seaward of the lagoon deposits are sandy, often cross-stratified marine deposits in which the most common fossil is a type of brachiopod that was adapted to such agitated conditions (Figure 5-23C).

In finer-grained sediments deposited farther offshore is a fossil community that includes many species, none very abundant. Many types of brachiopods are present, and trilobites are restricted to this belt. The high diversity of species reflects the stable conditions of an offshore shelf environment, one beyond the influence of river water and storm waves; the low abundance of species reflects a weak food supply, far from the algae and primitive plants that must have flourished in the vicinity of the lagoon and supplied its inhabitants with food. Where muddy sediments occur in this area, planktonic graptolites—fragile colonial animals that settled to the quiet seafloor after death—are preserved (Figure 5-23D).

Fossils have played a key role in the reconstruction of this set of Silurian environments. In the next section, we will examine deposits that are composed entirely of recognizable fossils and fossil debris.

**Organic reefs are bodies of carbonate rock**

In tropical shallow marine settings where siliciclastic sediments are in short supply, carbonate sedimentation usually prevails. Here coral reefs are often prominent,
Fossils found in Silurian rocks of Wales. These rocks contain fossils of marine invertebrates that would have been present in successive marine habitats from an inner lagoon rising above the seafloor as rigid structures. These organic reefs, which are produced largely by organisms that secrete calcium carbonate, form their own distinct depositional records—as bodies of limestone. Although some ancient reefs were formed by organisms other than corals, they, like their modern counterparts, grew in shallow waters of high clarity and normal marine salinity.

The basic framework of a reef consists of the calcareous skeletons of organisms, primarily corals. This framework is strengthened by cementing organisms that encrust the surface of the reef. Carbonate sediment, composed of fragments of the skeletons of reef-dwelling organisms, is trapped within the porous framework of the reef, filling some voids. With their complex internal structure, reef limestones are typically either unbedded or only poorly bedded. Even with the presence of infilling debris, reef limestone is so porous that many ancient buried reefs serve as traps for petroleum, which migrates into them from sediments rich in organic matter.

Because living reefs stand above the neighboring seafloor, they alter patterns of sedimentation nearby. On the leeward side of an elongate reef—the side nearest the land—there is often a relatively calm lagoon, especially if the reef has a typical reef flat, or horizontal upper surface, that stands close to sea level (Figure 5-24). Below the living surface of the reef is a limestone core consisting of a dead skeletal framework and trapped sediment. A pile of rubble called talus, which has fallen from the steep, wave-swept reef front, often extends seaward from the living surface.
Reefs build upward rapidly enough to remain near sea level even when the seafloor around them is becoming deeper. Many reefs, in fact, grow so rapidly and are so durable that they build seaward in the manner of a prograding delta. Figure 5-25 shows a spectacularly exposed cross section of a Devonian reef in Australia. Although it was built by organisms that have been extinct for hundreds of millions of years, this reef closely resembles many modern reefs in its basic structure—it displays both seaward talus deposits and leeward back-reef strata.

Isolated patch reefs are often found in lagoons behind elongate reefs (see Figure 5-24). Elongate reefs that face the open sea and have lagoons behind them are known as barrier reefs (see p. 103). Reefs that grow right along the coastline without a lagoon behind them are known as fringing reefs. Some fringing reefs grow seaward and eventually become barrier reefs.

Perhaps the most curious reefs in the modern world are the circular or horseshoe-shaped structures known as atolls. Atolls form on volcanic islands and thus are quite common in the tropical Pacific, which is dotted by many such islands. Charles Darwin's explanation for the origin of Pacific atolls is still accepted today (Figure 5-26). According to Darwin, each atoll was formed when a cone-shaped volcanic island was colonized by a fringing reef. The island then began to subside, turning the reef into a barrier reef, with a lagoon separating it from the remnant of the volcano. The island eventually sank beneath the sea, leaving a circular reef standing alone with a lagoon in the center, where limestone now accumulates in quiet water. Often the reef does not quite form a full circle, but is broken by a channel on the leeward side, where food supplies are low and reef-building organisms do not thrive. Horseshoe-shaped atolls range up to about 65 kilometers (40 miles) in diameter; during World War II their lagoons served as natural harbors for ships.

Ancient atolls that lie buried beneath younger sediments can be identified by studying cores of rock brought up from drilling operations—and because porous reef rocks often serve as traps for petroleum, drilling in the

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**FIGURE 5-25** Outcrop along Windjana Gorge, northwestern Australia, revealing the internal structure of a Devonian reef. The reef core consists of unbedded limestone. The talus is crudely bedded, and the beds slope away from the reef core.

The back-reef strata are also crudely bedded, but the beds are approximately horizontal. (Image courtesy of the Geological Survey of Western Australia, Department of Mines and Petroleum. © State of Western Australia 2013.)
FIGURE 5-26 Development of a typical coral atoll in the Pacific. The four stages of development of an atoll, as proposed by Charles Darwin. The photograph shows a coral reef encircling a volcanic island in the Pacific Ocean with a lagoon behind; it represents stage 3 of atoll formation. (Jean-Marc Truchet/The Image Bank/Getty Images.)

FIGURE 5-27 A late Paleozoic atoll. This horseshoe-shaped atoll lies almost a kilometer (0.6 mile) below the surface of the land in Texas. The atoll was discovered when rocks of the region were drilled for petroleum. The reef appears to have faced prevailing winds from the south. (After P. T. Stafford, U.S. Geological Survey Professional Paper no. 315-A, 1959.)

vicinity of these atolls is often profitable. Figure 5-27 shows the outline of a subsurface atoll of late Paleozoic age that has yielded considerable quantities of petroleum in the state of Texas. This atoll did not form on a volcanic island in the manner of the atolls studied by Darwin in the Pacific Ocean.

Carbonate platforms form in warm seas

An organic reef commonly forms part of a carbonate platform, which is a broad structure that consists of calcium carbonate and stands above the neighboring seafloor on at least one of its sides. Organic reefs often grow along the windward margins of carbonate platforms, where food is plentiful. The calcium carbonate of carbonate platforms precipitates from shallow tropical waters at or near the site where it accumulates.

Reefs and carbonate platforms are largely restricted to tropical seas because carbon dioxide is less soluble in warm water than in cold water. (You may have noticed that a warm bottle of carbonated soda fizzes readily when shaken, releasing bubbles of carbon dioxide.) Removal of carbon dioxide from seawater favors the precipitation of calcium carbonate because dissolved carbon dioxide combines with seawater to form carbonic acid,

$$\text{CO}_2 + \text{H}_2\text{O} = \text{H}_2\text{CO}_3$$

The cooling of water in contact with the atmosphere increases the concentration of dissolved carbon dioxide in that water, driving this chemical reaction to the right and increasing the concentration of carbonic acid. Similarly,
The Bahama banks, which are now separated from Florida by the Straits of Florida. Here great thicknesses of carbonate sediment have accumulated over 170 million years.

The varied sediments currently accumulating on the Bahama banks resemble those of many ancient carbonate platforms. Here, as on carbonate platforms generally, sediments accumulate rapidly. Since mid-Jurassic time, about 170 million years ago, some 10 kilometers (6 miles) of carbonates have accumulated both on the Bahama banks and in southern Florida, which was part of the same carbonate platform during Cretaceous time and earlier. This heavy buildup of sediments has caused the oceanic crust to subside, so that shallow-water Jurassic deposits now lie up to 10 kilometers below sea level.

Among the accumulating carbonate sediments are oolites, which are composed of ooids—spherical grains that consist of aragonite needles precipitated from seawater (see Figure 2-23). Strong currents pile ooids into spectacular shoals in some shallow subtidal areas of the Bahama banks (Figure 5-29). Oolites formed in this manner display conspicuous cross-bedding. This feature, and the need for individual ooids to roll around in order to form, makes cross-bedded oolites in the rock record reliable indicators of shallow seafloors swept by strong currents.

Within tidal channels in some areas of the Bahama banks are knobby intertidal structures known as stromatolites. They are produced by threadlike cyanobacteria.
FIGURE 5-29 Aerial view of oolite shoals along the Great Bahama Bank, with deep water on the left. These shoals are up to 8 kilometers (5 miles) long. (Paul M. [Mitch] Harris with permission of Chevron Petroleum Technology Company.)

As Figure 5-30 indicates, these organisms form sticky mats that trap carbonate mud. After forming a mat, they grow up through it to produce another one. Repetition of this process on an irregular surface forms a cluster of stromatolites. Each is internally layered: organic-rich layers alternate with organic-poor layers. The fossil record of stromatolites is unusually ancient, extending back nearly 3.5 billion years.

There is a simple reason that stromatolites are found almost exclusively in supratidal and high intertidal settings. Because these environments are exposed above sea level much of the time, they become hot and dry, and relatively few marine animals can survive in them. Thus there is little to interfere with the tendency of the mats created by cyanobacteria to form layered structures. Such mats can grow under water, but they are quickly eaten by grazing marine animals and damaged by burrowers, so they seldom accumulate to form stromatolites or well-layered limestones. Exceptions are the large, column-shaped stromatolites of the Bahamas, which grow in subtidal channels where tidal currents are very strong. Few animals can survive in these current-swept areas, so stromatolites flourish there. Stromatolites also flourish in Shark Bay, Western Australia, where the waters are hypersaline and animals are very rare (Figure 5-31). As we will see in later chapters, very ancient stromatolites formed beneath the sea before the origin of marine animals that feed on cyanobacteria.

FIGURE 5-30 The growth of stromatolites. A. A mat of cyanobacteria traps sediment, and the cyanobacteria grow through it to form another layer. The accumulation of several layers leads to the formation of a stromatolite. Stromatolites range from a few centimeters to a few meters in width. B. An example of a stromatolite from the Early Ordovician Prairie du Chien Group of eastern Wisconsin. This stromatolite was deposited in an intertidal environment, along with ooids (at the base of the sample). It has been subsequently replaced by dolomite. It is about 6 inches (16 centimeters) wide, or about the size of a small head of cabbage. (John Lucza, University of Wisconsin–Green Bay.)
they reach gentler slopes and spread out far from shore. In the 1930s the Dutch geologist Philip Kuenen demonstrated in the laboratory that turbidity currents can attain great speed, especially when they are heavily laden with sediment and moving down steep slopes. Sediment suspended in a turbidity current behaves as part of the moving fluid, and its presence increases the density of the fluid by as much as a factor of 2.

When the slope beneath a turbidity current begins to flatten out, the current slows and spreads out, dropping its sediment in the general sequence that we have seen again and again: first the coarse sediment falls from suspension, and then, much later, the fine material follows. The result is a graded bed of sediments, with poorly sorted sand and granules at its base and mud at the top. Such a graded bed is known as a turbidite (Figure 5-32A).

Deep-Sea Environments

To reconstruct the distribution of continents and ocean basins for any time interval in the past, geologists must identify not only nearshore deposits of the various kinds just described, but also deposits formed in deep-water environments beyond continental shelves. Coarse clastic deposits derived from a continental shelf accumulate along its base, but fine-grained sediments predominate in the middle of a huge ocean, far from sources of clastic sediments, and they accumulate very slowly.

Turbidity currents flow down submarine slopes

One of the most remarkable advances in sedimentology took place in the middle of the twentieth century with the recognition that certain sedimentary rocks have been produced by turbidity currents. A turbidity current is a flow of dense, sediment-charged water moving down a slope under the influence of gravity.

Turbidity currents were first noticed in clear lakes, where flows form from muddy river water that hags the lake floor. These currents flow for a considerable distance, slowing and dropping their sediment only when

Figure 5-32 Turbidites. A. A turbidite in carbonate rock from the middle Cretaceous of Mexico. The arrows show the top and bottom of the graded unit, which is about 15 centimeters (6 inches) thick. B. The bottom surface of a turbidite bed, showing “sole marks” produced when the sediment forming the bed filled depressions that formed immediately after a turbidity current scoured the preexisting sediment surface. These particular scour marks, known as flute casts, are about 5 centimeters (2 inches) wide. The scour marks become wider and thinner toward the lower left, indicating that the current flowed in that direction. (A. Courtesy of Paul Enos, University of Kansas; B. Richard Becker/FLPA/age fotostock.)
Large turbidity currents flow down continental slopes and deposit turbidites along continental rises and on the abyssal plain. Turbidity currents that originate near the edge of the continental shelf not only carry sediment to the deep sea, but also erode both the continental slope and part of the continental rise. Thus such currents are largely responsible for carving great submarine canyons in many parts of the slope. The turbidity currents slow down at the mouths of these canyons, dropping part of their sediment load to form deep-sea fans that superficially resemble alluvial fans (Figure 5-33). In fact, much of the continental rise actually consists of coalescing submarine fans.

Ancient turbidites are normally stacked one on top of another in groups. The result is that the deposits they form are cyclical (see Figure 1-21). The bottom of each cycle consists of coarse material at the base of the turbidite, and the top consists of mud that accumulated in the quiet depths before the next turbidite formed.

The sandy portions of lithified turbidites are typically graywackes, which are quite unlike the clean sands of meandering river deposits. This is only one of several ways in which a cyclical sequence of turbidites differs from a meandering-river cycle, although both show a grading of sediment from coarse to fine within each complete cycle. Another difference is that a single turbidite is usually only a few centimeters and seldom as much as a meter thick (see Figure 1-21), whereas most complete meandering-river cycles measure at least 2 or 3 meters (6-9 feet) from bottom to top. In addition, turbidites lack the large-scale cross-bedding characteristic of meandering river channels. Furthermore, the base of a turbidite is often irregular because the earliest, most rapidly moving waters of a turbidity current scour depressions in the sedimentary surface laid down earlier. These scour are subsequently filled in by the first sediments that settle as the current slows. When the base of a lithified turbidite is turned over for inspection, its irregularities, or "sole marks," can reveal the direction of water flow (Figure 5-32B).

**Pelagic sediments are fine-grained and accumulate slowly**

Turbidity currents and other bottom flows carry mud to the abyssal plain of the ocean well beyond the continental rise. None of these flows, however, contributes sediment to the deep sea at a sustained high rate; indeed, sediment in most areas of the abyssal plain accumulates at a rate of about a millimeter per thousand years! In the deep sea, most sediments are clays, which come from two sources. One source is the weathering of rocks produced by oceanic volcanoes. Such clays are less abundant in the Atlantic than in the central Pacific, where volcanoes are common. Clays also reach the deep sea by settling from the water above, having traveled through the air as wind-blow dust or having drifted seaward from the land at very low concentrations in surface waters of the ocean. Just as organisms that occupy the water above the deep-sea floor are termed pelagic forms of life, fine-grained sediment that settles from these waters to the deep-sea floor is called pelagic sediment.

In addition to clays, pelagic sediment includes skeletal material contributed by small pelagic organisms. Whether they predominate in any area of the deep sea depends on the extent to which they are diluted by the more rapid accumulation of biologically produced sediments. Some of the latter consist of calcium carbonate, while others consist of silica.

Where deposition of calcium carbonate predominates in the deep sea, its fine grain size has led oceanographers to refer to it as calcareous ooze (Figure 5-34). This sediment consists of skeletons of single-celled planktonic organisms, including planktonic foraminifera, which are amoeba-like protozoans of the group Rhiararia (see Figures 3-17 and 3-6). Other important constituents are the armorlike plates that surround coccolithophores, the single-celled floating algae that are major components of tropical phytoplankton (see Figure 3-16C).
Concentration of carbonic acid. As a result, most small particles of calcium carbonate dissolve by the time they sink to a depth of 4,000 meters.

In many regions at high latitudes, as well as in tropical Pacific regions characterized by strong upwelling, biologically produced siliceous ooze carpets the deep-sea floor. This sediment consists of the skeletons of two groups of organisms that thrive where upwelling supplies nutrients in abundance: diatoms, a highly productive phytoplankton group found in nontropical waters (see Figures 3-16B and 3-6), and radiolarians, which are single-celled planktonic protozoans related to foraminifera (see Figures 3-18 and 3-6). Recall that the skeletons of both diatoms and radiolarians consist of a soft form of silica called opal (see Figure 2-21), which tends to recrystallize so that individual skeletons cannot be discerned. In the process, the rock that they form becomes a dense, hard chert (see Figure 2-20). Before recrystallization, soft sediment, consisting largely of diatoms, is known as diatomaceous earth—the abrasive component of many scouring powders.

Thick bodies of diatomaceous sediment have formed in marine areas of strong upwelling. Diatoms did not exist until late in Mesozoic time, however, and here we come to an important point: the composition of pelagic sediments has changed markedly in the course of geologic time as groups of sediment-contributing organisms have waxed and waned within the pelagic realm.

**CHAPTER SUMMARY**

How does vertical stacking of distinctive types of strata provide clues to environments of deposition?

Sometimes one kind of rock alone serves to identify an ancient environment. Usually, however, suites of closely associated rock types are required for this purpose. These
deposits are commonly organized in sedimentary cycles in which, in accordance with Walther's law, one kind of sediment tends to lie above another that accumulated in an adjacent environment.

**What sedimentary features result from deposition in particular nonmarine environments?**

Ancient soils resembling those of the modern world are sometimes found beneath unconformities, although they may be hard to identify because of chemical alteration. Some types of ancient soils reflect the climatic conditions under which they formed. Lake deposits, which are much less common than marine deposits, are characterized by thin horizontal layers, few burrows, and an absence of marine fossils. Glaciers, which plow over the surface of the land, often leave a diagnostic suite of features, including scoured and scratched rock surfaces, poorly sorted gravelly sediment, and associated lake deposits that exhibit annual layers. In hot, arid basins, erosion of the surrounding highlands creates gravelly alluvial fans. Braided streams flowing from the fans toward the basin center deposit cross-bedded gravels and sands. Beyond these deposits may be shallow playa lakes and salt flats where evaporites accumulate. Some arid basins also contain dunes of clean, cross-bedded, windblown sand. In moist climates, braided streams also form on alluvial fans, and in lowland areas, meandering rivers leave characteristic deposits in which channel sands and gravels grade upward through point-bar sands to muddy floodplain sediments.

**What are the distinctive features of marginal marine and continental-shelf deposits?**

Where a river meets a lake or an ocean, it drops its sedimentary load to form a delta. Deltaic deposition typically produces an upward-coarsening sequence as shallow-water sands build out over deeper-water muds. More widespread than deltas along the margin of the ocean are muddy lagoons bounded by barrier islands formed of clean sand. Coral reefs border many tropical shorelines. A typical reef stands above the surrounding seafloor, growing close to sea level and leaving a quiet lagoon on its leeward side. Most reef limestones are supported by rigid internal organic frameworks. Coral reefs form parts of many carbonate platforms, although these platforms contain a number of other deposits as well. On continental shelves, storms occasionally produce thin sandy beds known as tempestites. Continental-shelf environments are usually dominated by either carbonate or siliciclastic sediments.

**What are the characteristics of deep-sea sediments?**

Beyond the edge of the continental shelf, turbidity currents intermittently sweep down continental slopes to the continental rise and abyssal plain, where they spread out, slow down, and deposit graded beds of sediment, known as turbidites. Still farther from continental shelves, only fine-grained pelagic sediments accumulate. Clay reaches these deep-sea areas very slowly. In some areas the deposition of clay is far surpassed by the accumulation of minute skeletons of planktonic organisms, which settle to the seafloor to form calcareous or siliceous oozes.

### REVIEW QUESTIONS

1. What kinds of nonmarine sedimentary deposits reflect arid environmental conditions?
2. What kinds of nonmarine sedimentary deposits reflect cold environmental conditions?
3. What kinds of deposits indicate the presence of rugged terrain in the vicinity of a nonmarine depositional basin?
4. In what nonmarine settings do gravelly sediments often accumulate?
5. Contrast the patterns of occurrence of sediments and sedimentary structures in the following three kinds of depositional cycles: the kind produced by meandering rivers, the kind produced by deltas, and the kind produced by turbidity currents.
6. Draw a profile of a barrier island–lagoon complex, and label the various depositional environments.
7. What features typify sediments that accumulate in the centers of lakes?
8. How do stromatolites form?
9. Describe the kind of rock found in a typical organic reef.
10. Where is a lagoon in relation to a barrier reef? Where is it in relation to an atoll?
11. Which features of carbonate rocks suggest intertidal or supratidal deposition? Which features suggest subtidal deposition?
12. What types of sediments and sedimentary structures usually reflect deposition in a deep-sea setting?
13. What are the important depositional environments of the basic kinds of sediment (such as mud, well-sorted sand, gravel, evaporites, and various kinds of limestone) and of particular sedimentary structures (such as cross-bedding, graded beds, mudcracks, and ripples)? How are the kinds of sediment and sedimentary structures found within each environment related to processes operating within it? Use the Visual Overview on pages 104–105 for reference.