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Marine Sedimentation

Nothing under heaven is softer
or more yielding than water,
but when it attacks things hard
and resistant there is not one
of them that can prevail.

—Tao Te Ching, sixth century BCE

preview

Much of the ocean floor is covered by layers of sediment; each has a story to tell about the geologic past. These sedimentary layers contain a great variety of particles. Some of these grains were swept to the ocean bottom from land, or dropped down from outer space, or were produced by a variety of biological or chemical processes. A careful study of the properties of these sedimentary deposits by trained geologists yields valuable clues about the Earth's plate-tectonic history, the evolution of marine life, the chronicle of past climates, variations in the flow pattern of water currents, impacts of meteorites, eruptions of submarine volcanoes, fluctuations of sea level, occurrences of mass extinction, and many more events. Epic stories of the Earth's past can be read from the record that is preserved in the vast sedimentary accumulations on the sea bottom.

The central topics of this chapter are the nature and significance of the tremendous amount of sediment that covers the floor of the deep sea and the submerged edges of the continents. As the oceans spread apart and the continents are dragged along, water wears down the bedrock of the land. We will examine the character and history of sedimentary material—rock debris and fossil remains—that is slowly being deposited in the ocean basins. These beds of sediment, when sampled and examined carefully, reveal clues about the chemistry and biology of ocean water of the distant past, the geologic history of ocean basins, the evolution and mass extinction of marine life, and the nature of past climates.

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4-1 Sediment in the Sea

Sediment is produced by the weathering (the chemical and mechanical breaking down) of rock such as granite or basalt into particles that are then moved by air, water, or ice. Sediment can also form by the accumulation of shells of dead organisms. Therefore, sediment can consist of either mineral or fossil particles, and mixtures of both types can be found in many places on the bottom of the sea. Most erosion of rock occurs on land, and most deposition of sediment occurs in the ocean. The net effect of these two processes—erosion and deposition—would be to even out the Earth's surface, cutting away its high points (the landmasses) and filling in its low points (the oceans), if it were not for plate tectonics. The collision of continents along subduction zones squeezes and deforms the layers of marine sediment that have accumulated between landmasses, raising long, linear mountain belts. The mountain peaks rise high above sea level until erosion wears them away, and the weathered debris is swept into the ocean and accumulates as sediment. This results over geologic time in a grand tectonic cycle of raising and leveling mountains and deepening and filling ocean basins. Even on the land, marine sedimentary rocks are common because seas have regularly invaded the land in the geologic past. These marine sedimentary rocks make up more than 50 percent of the outcrops on land and record the existence of extensive seas that dried up long ago.

Before examining some of the factors that determine the patterns and depositional rates of marine sediment, we need to be able to distinguish among different types of deposits. So let's consider two simple schemes that geologists use to do this.

Classification of Marine Sediment

The first step in classifying anything is to establish criteria for defining categories. Sediments can be subdivided on the basis of the size of the particles (**grain size**) or on the basis of their mode of formation. In the first case, the classification depends on a measurement of particle size; in the second case, the classification requires an interpretation of the origin of the deposit. Both classifications are useful and are widely employed by geological oceanographers. Let's examine both of them more carefully.

The size of particles produced from the breakdown of rock ranges from enormous boulders to tiny grains of microscopic clay or even finer particles called **colloids**.

TABLE 4–1

WENTWORTH GRAIN-SIZE SCALE

Sediment	Type	Diameter (mm)
Gravel	Boulder	>256.0
	Cobble	65.0–256.0
	Pebble	4.0–64.0
	Granule	2.0–4.0
Sand	Very coarse	1.0–2.0
	Coarse	0.50–1.0
	Medium	0.25–0.50
	Fine	0.125–0.25
	Very fine	0.0625–0.125
Mud Silt		0.0039–0.0625
Clay		0.0002–0.0039
Colloid		<0.0002

Source: Adapted from Wentworth, C. K. A scale of grade and class terms for classic sediments, *Journal of Geology* 30 (1922): 377–392.

From the largest to the smallest particles common in sediment, there are gravel, sand, silt, and clay (TABLE 4–1). We can ignore the even finer colloids because they are unimportant as sediment. Silt and clay particles are typically mixed together and form a deposit of **mud**. The most common sedimentary deposits in the ocean are mud and sand; gravel is very rare in the sea.

Now let's classify sediments by the way that they form. Using the broadest scheme possible, we can subdivide sediment into five categories, presented here with their general characteristics:

Terrigenous sediment: *fine and coarse grains that are produced by the weathering and erosion of rocks on land; typically sands and mud.*

Biogenous sediment: *fine and coarse grains that are derived from the hard parts of organisms, such as shells and skeletal debris: typically lime (composed of calcium carbonate) and siliceous (composed of silica) muds.*

Hydrogenous sediment: *particles that are precipitated by chemical or biochemical reactions in seawater near the seafloor; manganese and phosphate nodules are examples.*

Volcanogenous sediment: *particles that are ejected from volcanoes; ash is an example.*

Cosmogenous sediment: *very tiny grains that originate from outer space and tend to be mixed into terrigenous and biogenic sediment.*

Notice how the two classification schemes—one based on grain size, the other on sediment origin—are intertwined. For example, sand and mud, which are separated on the basis of grain size, can be terrigenous, biogenous, cosmogenous, hydrogenous, or

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volcanogenous, depending on their origin. Now that we have a common vocabulary for distinguishing types of sediment, let's consider how exactly the grains are transported and deposited on the seafloor to form those various types.

Factors that Control Sedimentation

Two of the most important factors that determine the nature of a sediment deposit are particle-size distribution and energy conditions at the site of deposition. Because these two factors interact to produce the properties of a sediment deposit, a geologist is able to deduce what they were at the time of deposition by examining such properties of even an ancient sediment.

Terrigenous sediment is a collection of rock and mineral fragments with compositions that are directly related to their source rocks. For example, if fragments of granite are present, then the source area that yielded the grains contained granitic rocks. If limestone and shale fragments abound in the deposit, then the source rocks were composed of sedimentary rocks, including limestones and shales. Rivers, glaciers, and wind transport these particles out of the source area, a process termed **erosion**. Much of this eroded sediment eventually reaches the ocean, where currents disperse them even more before they are deposited as layers of sediment on the sea bottom.

If the rock eroded slowly, then sediment was supplied at a low rate and likely was reworked by currents before it was buried by ever younger deposits. A slow rate of sedimentation generally gives more opportunity for water to sort the grains according to size, shape, and density. This can result in deposits of mud and sand that are well sorted (a small range of grain sizes) and have a uniform appearance, such as a coarse sand or very fine sand or silt. Rapid erosion supplies sediment at a high rate. Currents have little time to sort the grains before they are buried by still more sediment. As a result, such deposits tend to be poorly sorted (a broad range of grain sizes) and heterogeneous (nonuniform in appearance, such as gravel mixed with sand, or sand mixed with mud).

In most cases, a clear relationship exists between the average grain size and the strength (energy) of bottom currents at the time that sediment is deposited. Stated simply, the average particle size of a deposit is proportional to the energy level present at the time of deposition. Under high-energy conditions, water is swift and turbulent, keeping fine grains in suspension and resuspending those fine particles that momentarily settled to the ocean floor. This constant agitation of the sea bottom separates small grains and transports them into quieter water, which typically is deeper than turbulent water. Thus, a coarse sand (rather than medium or fine) is deposited under high-energy conditions. Low-energy environments, where currents are weak and water is quiet, do not receive a supply of coarse grains because the weak currents cannot transport them to these sites. Here muds typically accumulate. Consequently, the average grain size of a deposit of sediment serves as a good measure of the energy of the environment at the time of deposition. Fine-grained sediments denote low-energy conditions; coarse sediments, high-energy conditions.

In laboratory experiments, geologists have worked out the relationship between grain size and current velocity that specifies whether sediment of a particular size will be eroded, transported, or deposited under a given set of energy conditions. The general results of these experiments are summarized in the Hjulström diagram (FIGURE 4–1). This chart plots the average current velocity (the *y*-axis) against the particle diameter (the *x*-axis). Both axes use a log-to-the-base

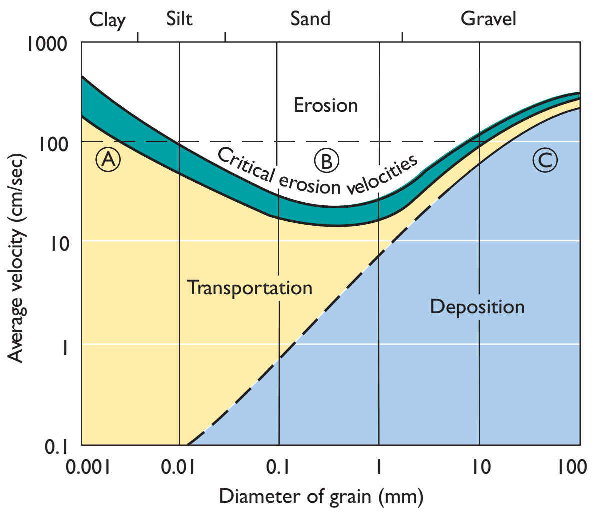

FIGURE 4–1 **Hjulström's diagram.** This plot shows the average current velocities necessary for the erosion, transportation, and deposition of sediment particles of different sizes. The broad upper curve indicates that strong currents are needed to erode mud (clay and silt), coarse sand, and gravel, whereas weaker currents are sufficient to erode fine to medium sand. The lower curve indicates that once sediment is moving, the current velocity necessary to transport grains varies directly with particle size. For a given grain diameter, deposition occurs when the current speed is less than some critical value defined on the graph by the lower line. The letters A, B, and C near the top of the chart are discussed in the text. [Adapted from Sundborg, A. *Geografiska Annaler* 38 (1956): 125–316.]

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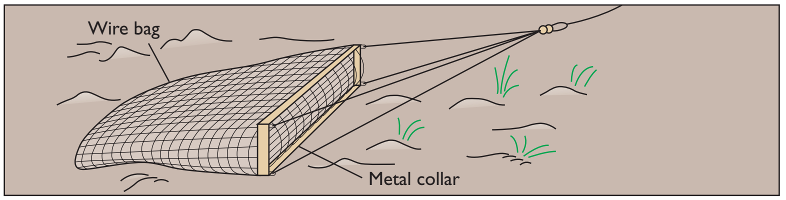
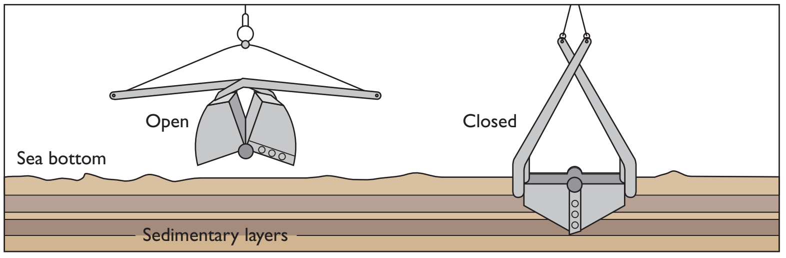
Geology

Probing the Seafloor

Various simple but durable devices are available for collecting sediment samples from the ocean floor, even from the deepest, most inaccessible parts of the sea. A long-established technique is scraping the ocean bottom with a **dredge**—a rigid metal frame to which is attached a sampling bag made of chain or tough netting (FIGURE B4–1a). Dredges are suitable for obtaining large, bulk samples of either rock or sediment. As they are dragged, however, they bite the bottom indiscriminately and mix samples together in the sampling bag. Also, fine sediment such as mud tends to be washed out of the sample. Because of these effects, oceanographers employ



(a) BOTTOM DREDGE



(b) GRAB SAMPLER

(c) GRAB SAMPLER

FIGURE B4–1 **Gear for sampling sediment.** (a) Hard rock can be sampled with the durable bottom dredge. (b) Surface sediments are collected by grab samplers that take a "bite" out of the sea bottom. (c) A grab sampler being lowered to the sea bottom.

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dredges almost exclusively to collect hard rock rather than soft sediment. Less disturbed samples of mud and sand are collected by **grab samplers**—spring-loaded metal jaws that take a bite out of the bottom and close tightly around the sediment sample (FIGURES B4–1b and 1c).

Dredges and grab samplers merely sample the surface veneer of sediment. Deeper penetration of soft sediment is accomplished by a **gravity corer**. This hollow metal tube, known as a **core barrel**, is pushed into the sediment by the force of gravity. The corer is lowered to the bottom, where the heavy weight at the top of the device drives the barrel into the sediment (FIGURE B4–2a). A plastic liner that has been inserted into the core barrel allows oceanographers to extract the sediment core intact from the sampler and also serves as a temporary storage container. Gravity corers are capable of taking cores of between 1 and 2 meters (~3 and 6 feet) long, depending on the properties of the sediment. Sediment cores longer than 20 meters (~66 feet) are routinely obtained by **piston corers** (FIGURE B4–2b). This type of corer has a piston that slides up the core barrel as it penetrates the bottom. The action of the piston extrudes water from the core barrel, allowing the sediment core to enter the liner with minimal disturbance and compaction. Once the core is on deck, the plastic liner with its sample of sediment is extruded from the core barrel and taken to a laboratory for detailed examination (FIGURE B4–3). Geologists carefully study the layering and composition of the sediment particles to determine the geological history of the Earth.

At present, the best technique for sampling the ocean bottom is **platform drilling**, which was first developed by petroleum engineers on land and is now adapted to the ocean, even the deep ocean. The procedure is very expensive, but the scientific results are priceless. Marine geologists not only recover cores of sediment more than 1 kilometer (~0.6 miles) in length, but also can drill into the hard rock of the crust beneath the sedimentary layers. The *Glomar Challenger*—the 122-meter-long (~403 feet) vessel that completed an illustrious international career of drilling, even in the remotest regions of the oceans, collecting hundreds of kilometers of core sample—has been retired. Its successor, the larger *Joides Resolution*, has a much more efficient and deeper drilling capacity than the *Challenger* (see Figure 1–10 in Chapter 1).

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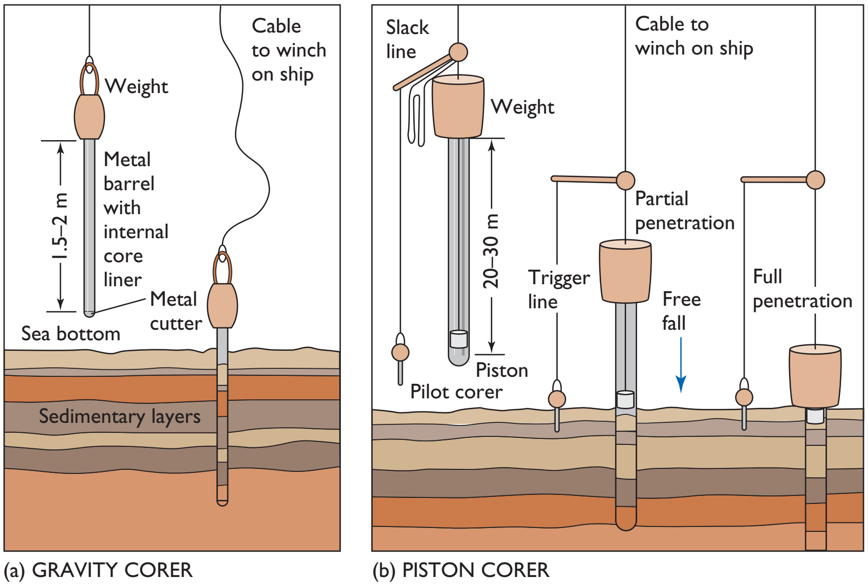

FIGURE B4–2 **Corers.** (a) A heavy weight drives the core barrel into the sediment. (b) A piston corer can take a much longer core than can a gravity corer because of a piston in the core barrel.

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FIGURE B4–3 A geologist on board the Japanese drilling ship *Chikyu* examines variations in the size and composition of sediment particles in a core of deep-sea layers in order to interpret ancient environments and climatic history.

10 (0.1, 1, 10, 100). Two curves divide the area of the plot into three distinct fields; these are labeled erosion, transportation, and deposition. The top curve specifies the speed that a current must have in order to erode grains of a particular size from the bottom. Note that for sand and gravel, the larger the particle, the stronger the current must be to erode the material. Surprisingly, though, higher current velocities are needed to erode tiny clay and silt particles than to erode fine sand! At first thought, this may not appear to make sense. Part of the explanation lies in the fact that mud—a mixture of clay and silt—is more cohesive and, thus, "stickier" than sand, and so a stronger current is needed to dislodge a mud particle from the bottom than a fine grain of sand. However, if you examine the lower curve on the diagram, it shows that once eroded and moving, a clay particle, but not a fine particle of sand, can be transported by a weak current. This is because the settling rate of suspended particles also varies with their diameter. Small particles settle more slowly than do large grains and so are kept in suspension longer and are easily moved about by weak currents.

Let's review these important ideas before proceeding. Look at the top of the Hjulström diagram. The top line represents a current velocity of 1,000 centimeters per second (~22 mph). Note that this strong bottom current will erode all grain sizes ranging from clay-size to gravel-size material. Now examine the 100 cm/sec line (~2 mph). Under this energy condition, clay (Point A) will be transported but not eroded, grains ranging in size from silt to fine gravel (Point B) will be eroded, and coarse gravel (Point C) will be deposited. A bottom current flowing at 10 centimeters per second (~0.2 mph) is too weak to erode any sediment at all. Yet these current speeds are capable of transporting clay, silt, and sand, provided they are already in suspension, but not gravel, which will be deposited.

4-2 Sedimentation in the Ocean

Away from a continent, two things happen: the water gets deeper and the sea bottom is farther from the source of terrigenous sediment, which is mainly sand

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and mud eroded from the land and supplied by rivers. On the basis of water depth, the ocean can be subdivided into two major areas of sedimentation. The continental shelf is shallow and close to sources of sediment from the land, and the deep sea is deep and far from river supplies of terrigenous sediment. Sedimentation processes are quite different in these two areas, and therefore each is examined separately.

Shelf Sedimentation

A continental shelf is a relatively broad, essentially flat platform 70 to 100 kilometers (~43 to 62 miles) wide that represents the submerged edge of a continent. Water depths on the shelf are shallow, varying from 0 at the shoreline to 120 to 150 meters (~396 to 495 feet) at the shelf break, where the gradient of the sea bottom steepens to about 4 degrees and marks the beginning of the continental slope. The seafloor of the shelf is nearly horizontal, with a regional slope that rarely exceeds 1 degree. Energy for eroding and transporting sediment grains is provided by the tides and wind-generated waves and currents.

Over most continental shelves, waves seem to be the dominant process affecting the sea bottom. Drawing from our swimming experience at the beach, we know that large waves contain more energy than do small waves. Diving beneath an unbroken wave, we will notice that the water becomes increasingly calmer with depth. In fact, if we go deep enough, the water will be relatively calm despite the surface agitation by waves. The bigger the waves, the deeper we must dive in order to escape the wave motion. Thus, we can infer that *bottom energy* induced by surface-water waves must diminish with distance offshore because water depths increase seaward.

The shoreline is affected by breaking waves, and these high-energy conditions suspend and remove all the fine sediment and allow mainly medium and coarse sand and gravel to be deposited on the beach and in the nearshore zone. Seaward from the nearshore zone, bottom energy induced by waves decreases because of increasing water depths. This decrease of bottom energy with water depth results in a systematic decrease in grain size with distance offshore. The beach, which is composed of coarse to medium sand and gravel, grades into fine sand farther offshore and, moving seaward, into muddy sand (sand with some mud), sandy mud (mud with some sand), and finally mud far offshore (FIGURE 4–2a).

There would be a decrease in grain size as the water depth over the shelf increased if sea level remained relatively fixed over time. Yet we know that because of glaciation and deglaciation in the recent geologic past, sea level has gone up and down. Sometimes it flooded the shelf as it does at present, and at other times it exposed the shelf so that it became covered with grasslands and forests rather than seawater. Sea level about 15,000 years ago was 130 meters (~429 feet) lower than it now is (FIGURE 4–2b). At that time the shoreline was located several hundred kilometers farther seaward than it is at present, virtually at the shelf break (FIGURE 4–3a). Consequently, beaches in the past were located some 150 kilometers (~93 miles) seaward of New York City (see Figure 4–3a) and about 300 kilometers (~186 miles) southeast of Texas. Since that time, glaciers have been melting, causing the seas to rise and flood the continental shelf. As the shoreline has advanced over the land, coarse sand and even gravel have been deposited on the outer shelf because of the shallow water depths that existed there in the past. Sea level is still rising worldwide, and much more flooding of land will occur if the ice sheets continue to melt.

So it seems that our original premise that grain size varies systematically with water depth is a valid one. Shallow waters over the inner shelf tend to be high-energy environments where coarse sediment accumulates, whereas deep waters over the outer shelf tend to be low-energy environments where fine sediment is deposited. However, we must take into account fluctuations of sea level that cause the water depths at any point on the continental shelf to vary as a function of time. This explains why coarse sediment (sand and even gravel) blankets the outer shelf where water depths are deep and the bottom is quiet. These coarse-grained deposits on the outer two-thirds of the shelf are not in equilibrium with the low-energy conditions that exist there at the moment. Such material is termed **relict sediment** because it accumulated at an earlier time and under very different depositional conditions. By contrast, the coarse to fine sands that cover the inner third of the continental shelf are modern sediments that are in equilibrium with the bottom energy conditions. There has not yet been sufficient time for fine sediment to bypass this inner band of modern sand and be deposited farther offshore and cover the relict sediment (FIGURE 4–3b).

Worldwide Distribution of Shelf Sediments

Many surveys of the distribution of sediment types of the continental shelf have been conducted worldwide. What emerges from these studies is a regular pattern of sediment types that vary with latitude (FIGURE 4–4a)

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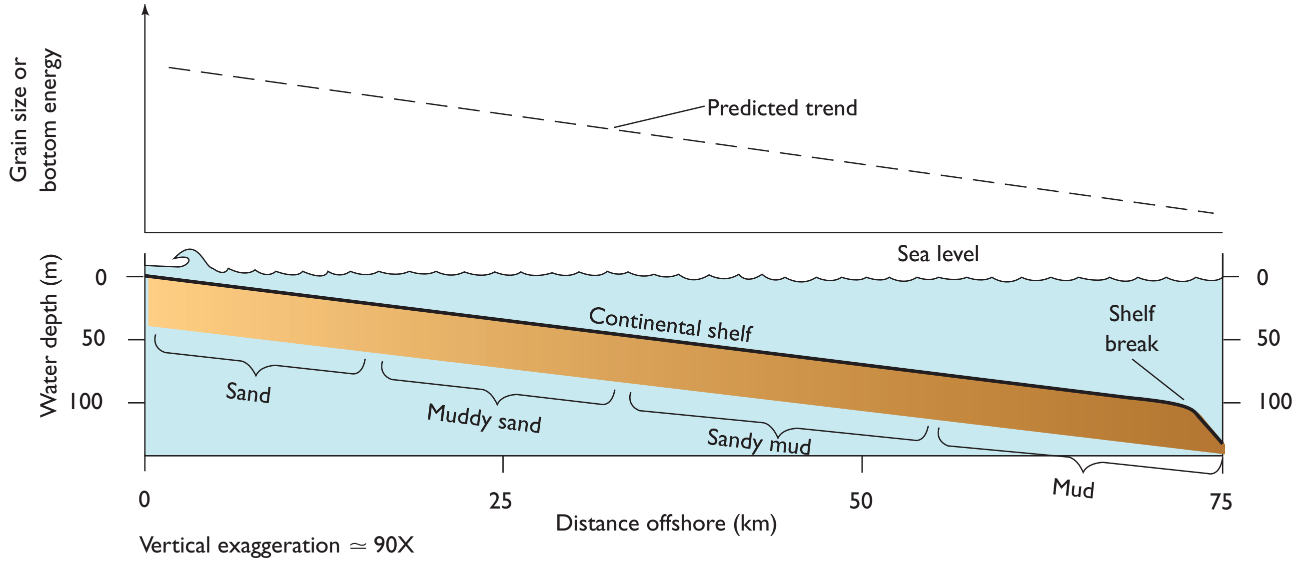
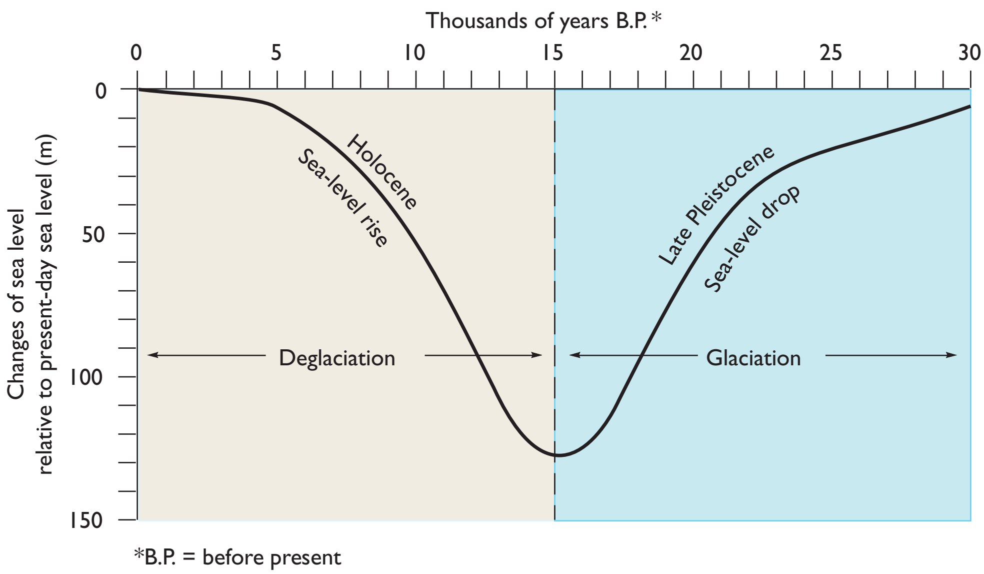


(a) MODEL PREDICTION OF SHELF SEDIMENTS



(b) POSITION OF SEA LEVEL FOR PAST 30,000 YEARS

FIGURE 4–2 **Shelf sedimentation.** (a) Assuming that bottom energy on the continental shelf is inversely proportional to water depth and thus distance offshore, the sedimentary cover should grade systematically from coarse sands and gravel onshore to mud at the shelf break. (b) This curve shows that sea level was 130 to 140 meters below its present level some 15,000 years ago. Since then, sea level has risen to its current position, an event known as Holocene sea-level rise. [Adapted from Emery, K. O. *Sci Am*. 221 (1969): 106–122.]

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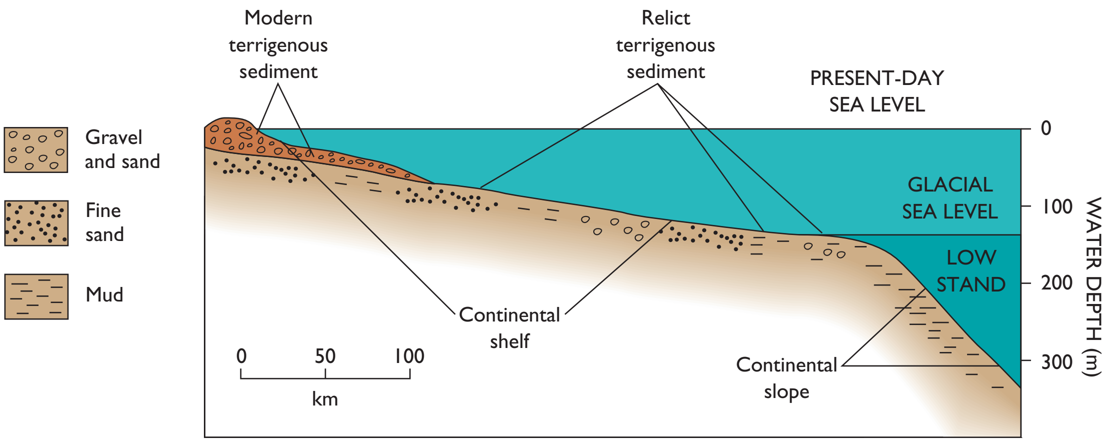
(a) COASTLINES PAST AND FUTURE



(b) RELICT SEDIMENT

FIGURE 4–3 **Shelf sedimentation.** (a) About 15,000 years ago, the shorelines were displaced seaward by more than 100 kilometers, as sea level dropped due to widespread glaciation. Presently, sea level is rising worldwide. If the world's ice sheets continue to melt, the oceans will flood many of the world's major cities in the future, as shown by the dotted line on the map of the East Coast of the United States. [Adapted from Emery, K. O. *Sci Am*. 221 (1969): 106–122.] (b) This cross section shows that modern terrigenous sediment forms an apron over the inner continental shelf and that most deposits are relict in nature, having been deposited during one or more previous low stands of sea level.

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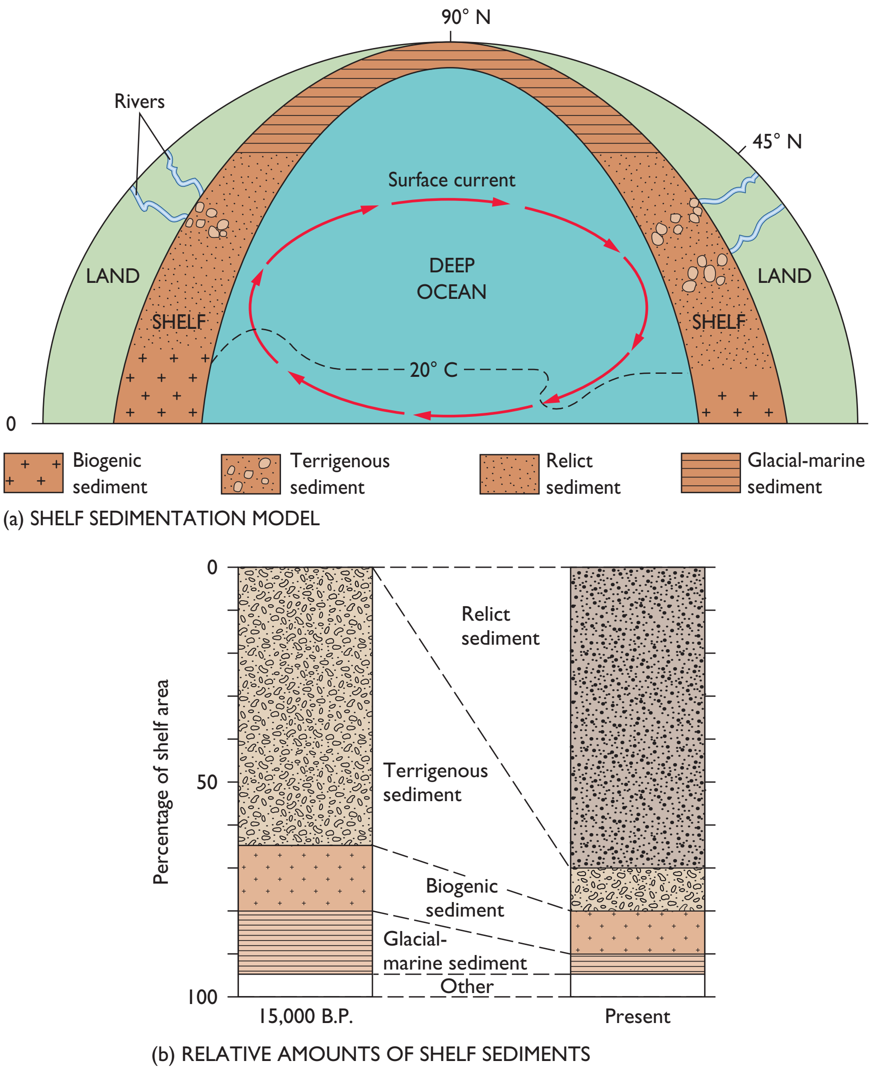

FIGURE 4–4 **Distribution of shelf deposits.** (a) To a first approximation, sediments on the continental shelves of the world vary with latitude. Biogenic sediment dominates the low latitudes, terrigenous sediment the middle latitudes, and glacial-marine sediment the polar regions. [Adapted from Emery, K. O. *Sci Am*. 221 (1969): 106–122.] (b) About 60 percent of the sediment that blankets the world's continental shelves is relict in nature, meaning it was deposited in the past under conditions that are no longer evident. Relict sediments are not in equilibrium with the present-day shelf environments. [Adapted from Emery, K. O. *Sci Am*. 221 (1969): 106–122.]

and that depend on climate. A broad band of biogenic sediment straddles the equator and extends into the subtropics. These deposits include coral reefs and accumulations of grain fragments composed principally of calcium carbonate ($CaCO_3$) derived from the hard parts of organisms such as clams, snails, sand dollars, coral, and calcareous algae. This band is broader along the western than the eastern edges of the oceans. Warm, westward flowing equatorial currents diverge from the equator and move poleward along the western sides of all the ocean basins. In contrast, the eastern sides of the ocean basins are bathed by cold currents that originate from high latitudes (see Figure 4–4a). Therefore, coral reefs that require warm water for growth occupy a broader part of the shelf along the western than along the eastern edges of ocean basins. The continental shelves of the middle or temperate latitudes are covered by river-supplied terrigenous deposits, principally sand-sized grains of quartz and feldspar derived from the weathering of granite on land. The polar shelves, not surprisingly, are littered with poorly sorted glacial deposits, either glacial **till** (unsorted deposits of boulders, gravel, sand, and mud) dumped there directly by glaciers or ice-rafted debris dropped from melting icebergs.

Recall, however, that most of the sedimentary cover of the continental shelves is not of recent origin but is relict in nature. This material is not currently

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in equilibrium with the present-day water depths (or energy levels). Rather, it was deposited long ago when shorelines were displaced seaward of their present positions. The recent **Holocene sea-level rise** (see Figure 4–2b) has been so rapid in a geologic sense that shelf sediments have not had sufficient time to regain equilibrium with the new, deeper water conditions. No more than 30 to 40 percent of the sediment on the surface of the continental shelf is recent (modern) in origin, and this "new" sediment is confined largely to the inner shelf. The remaining 60 to 70 percent is relict in character (FIGURE 4–4b).

Geologic Controls of Continental Shelf Sedimentation

In order to understand the geology of continental shelves, it is necessary to recognize the importance of time in their development. Asking which factors distribute sediment across the shelf from day to day is very different from posing the same question for longer time spans, such as millennia to millennia (thousands of years) or geologic period to geologic period (tens of millions of years). An oceanographer studying sedimentation on the Gulf Coast shelf of Texas during the past year will draw conclusions far different from those of a geologist investigating the million-year-long sedimentary record of the same shelf. To clarify this very important point, let's assume that this part of the Gulf Coast is ravaged by hurricanes on the average of once every fifty years or so. Chances are very good that no hurricanes would have occurred during the brief one-year field survey. About 20,000 hurricanes would, however, have swept through the area in the past million years. Thus, what is perceived as a rare storm event when measured against a human lifetime becomes a regular, ongoing geologic process of critical importance to an understanding of the million-year-long sedimentary history of the area. Given this distinction, we will consider the geologic development of continental shelves from the perspective of three very different time frames: a thousand years to the present day, a million (10^6) to a thousand (10^3) years, and 100 million (10^8) to a million (10^6) years. (Before continuing, turn to Appendix III and take a minute to study the geologic time scale.)

▪ **10^3 TO 0 YEARS:** The comparatively brief time interval of a thousand years to the present day (10^3 to 0 years) focuses on the "day-to-day" sedimentation processes on the continental shelves. Ocean surveys conducted during the past few decades probably are quite representative of conditions in general for the past thousand years. The climate has been relatively stable, and sea level has risen slightly—about a meter (~1 yard) or so—during this time span. Numerous ocean studies worldwide indicate that various currents move sediment across the shelf. The principal currents are generated by winds and tides.

Let's consider *wind effects* first. As the wind blows across the ocean's surface, it creates waves and currents. Waves grow in size and acquire more energy as wind velocity intensifies. Under calm winds, waves are small and have little effect on the shelf bottom. Storm waves, in contrast, are large and packed with energy, and their motion can extend downward to the very deepest parts of the shelf. As you will learn in Chapter 7, the water motion at the sea bottom induced by waves is back and forth, so that in theory there is no net movement of sediment grains. What waves do, however, is raise sediment particles off the bottom momentarily. Once in suspension, the particles are moved by other currents—even weak ones that, acting alone, could not erode the particle from the seafloor. For example, along the coasts of Washington and Oregon, marine geologists estimate that sediment on the continental shelf at a water depth of 80 meters (~264 feet) is moved by wave action about 10 percent of the year, mostly during winter storms. Sediment on the outer New England shelf also is affected largely during winter storms called "nor'easters."

The drag of the wind on the water surface, if persistent, also produces currents that deepen with time. Because these currents are weather related, the direction and strength of their flow vary markedly from day to day. Under fair-weather conditions, wind-generated currents are weak and exert little influence on the shelf bottom. Currents raised by powerful gale winds, however, can attain impressive speeds and move even coarse sand on the shelf bottom. For example, wind-generated currents moving faster than 80 centimeters per second (~1.7 mph) transport silt and sand (see Figure 4–1) across the Washington–Oregon shelf.

Now let's consider *tidal currents*. The water level on continental shelves fluctuates daily in response to the tides. Locally, the strength of the tidal currents depends on the tidal range (the vertical distance between high and low tides), the slope and roughness of the sea bottom, and the configuration of the shoreline. Commonly, tidal currents by themselves are too weak to rework bottom sediment. Some shelves, however, are dominated by tidal *scour* (sediment-bed erosion), especially those that are affected by a large tidal range and have tidal currents that are constricted by land or by a shallow bottom. The effect is something like what happens when you place your thumb over the spigot of a hose that is dribbling water. The constriction of the flow by your thumb causes it to shoot out strongly. A

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prime example of this effect is the shallow shelf around the shores of England, France, and Ireland. Here the speed of tidal currents regularly exceeds 100 centimeters per second (~2.2 mph) and locally exceeds 500 (~11 mph). An inspection of the Hjulström curve (see Figure 4–1) shows that such currents are capable of transporting sediment as coarse as gravel.

In contrast to wave-dominated shelves, which are static most of the year because of fair weather, much of the sediment of tide-dominated shelves is mobilized each day by bottom currents. However, the greatest volumes of sand are still moved during storm activity. Under storm conditions, the tide- and wave-generated bottom currents act together, causing an enormous drift of sand and gravel along the sea bottom. It has been estimated that the transport rate of sediment on the tide-dominated shelf of England is increased substantially during a storm, perhaps by as much as a factor of 20.

▪ 10^6 TO 10^3 YEARS: By geologic standards, a million (10^6) years is a mere fleeting moment of time. What occurs over the course of a million years usually has little consequence for the overall evolutionary history of a continental margin, which can stretch over hundreds of millions of years. An exception to this way of thinking, however, is the record of glaciation and deglaciation events over the past few million years. These are so current that their record has not yet been destroyed by more recent events and driven into geologic obscurity. This is like trying to remember details of your life as a six-year-old child when you are sixty years old. There is not much that you can recall about those times. Yet when you turned seven, abundant details of the previous year were fresh in your mind, not yet having been obliterated by living a long life.

The last two million years (the **Pleistocene epoch** or Ice Age; see Appendix III) have been dominated by glaciation. During this time ice sheets affected the surface morphology and the sediment composition of the world's continental shelves by controlling the level of the oceans. Sea level worldwide went up and down in direct response to the expansion and contraction of ice caps. With the onset of glaciation, snow was stored on land and converted to glacial ice. This caused sea level to drop about 100 to 150 meters (~330 to 495 feet) below its present level, and the shelf surface became exposed land. During the warm interglacial intervals of the Pleistocene epoch, ice sheets melted, causing sea level to rise worldwide and flood the continental shelves. The Earth's temperature trends for the past thousand years of the current interglacial stage are examined in the featured box "Climate Variability and Change."

During the low stands of sea level, which were times when water was stored as ice on land, sedimentation on the shelves was altered in many ways. Some of the more important modifications follow.

1. Ice caps and glaciers extended onto the shelf proper, particularly in the high and middle latitudes. Here, moving ice scoured, abraded, and plucked the shelf bottom. In some cases, it smoothed out the tops of hills; in others, it gouged out holes and created considerable topographic relief. Also, glaciers dumped debris on the shelf floor, forming blankets, mounds, and ridges of glacial till.
2. As sea level dropped, rivers in the unglaciated middle and low latitudes extended their channels across the continental shelves, cutting into the marine deposits. This resulted in the deposition of river sands and mud over marine deposits. During the height of glaciation, river deltas formed at the shelf break, which was the location of the new shoreline at that time. Eventually, the exposed shelf surface became vegetated, supporting forests and grasslands, and was populated by terrestrial animals.
3. Deposition and erosion rates increased in the deep-sea environments that are adjacent to the continental shelves. During low stands of sea level, rivers dumped their sediment loads near the shelf break and on the upper continental slope. Many of these deposits were unstable and slumped, producing sediment-laden bottom currents, which flowed downslope into the sea because they were denser than the surrounding water. These currents, in particular, helped to carve out or deepen submarine canyons. Massive amounts of sediment were funneled through these submarine canyons onto the deep-sea floor. Many of these canyons are now less active because the shoreline and rivers with their large supply of sediment were displaced landward away from the heads of submarine canyons as sea level rose.
4. Coral reefs were killed on a massive scale. Reef banks that flourished during high stands of sea level were stranded high and dry once sea level dropped with the onset of a period of glaciation. This led to widespread death of coral reefs. New coral banks were established seaward of the old reef mass, or new coral growth occurred along the lower fringes of the older, now dead reef bank. When sea level rose during the warm interglacial periods, the reefs responded by growing upward and keeping pace with the rise of sea level.

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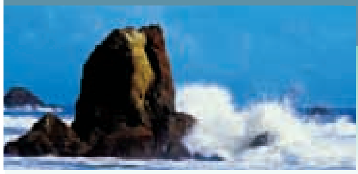

Geology

Dust Storms

In the Caribbean Sea, you can walk the deck of your boat anchored off the Virgin Islands and have your bare feet get covered with red African soil. It's hard to imagine that these small particles of sediment that have rained down on your vessel during the calm night originated from a massive dust storm that occurred a week before in the Sahara Desert of North Africa. Satellite images (FIGURE B4–4a) reveal that dust plumes intermittently cover enormous tracts of the North Atlantic Ocean. The photographs clearly show that these vast murky clouds originate in northern Africa, the bigger dust billows extending far to the west and reaching the southeastern United States, the Caribbean Sea, and even the far-away Amazon rainforests (FIGURE B4–4b). Worldwide, scientists estimate that perhaps as much as two billion metric tons of dust are blown into the air every year. Surprisingly, microorganisms and chemical toxins that adhere to these small particles are carried aloft as well and are dispersed far downwind from their points of origin. This flux of airborne particles indicates how the land, the ocean, the atmosphere, and, as we shall see, human activity are linked in a surprising variety of ways.

Actually, dust storms have been an ongoing presence in the region for millions of years, as indicated by the content of desert sand mixed into the pelagic mud that has been cored from the deep-sea bed of the North Atlantic Ocean. Recent investigations indicate that about 500 million metric tons of airborne dust advected from North Africa settles into the North Atlantic Ocean each year. During the last few decades, however, the number, size, and concentration of particles in these dust plumes have increased dramatically because of the agricultural and industrial impact of human beings living in northern Africa. Here, particularly in the Sahel region, which borders the southern fringe of the Sahara Desert, intensive farming to feed hungry and starving people combined with prolonged droughts that began in the 1960s has allowed desert winds to erode soil, microbes, pesticides, plastics, and other natural and industrial chemicals from northern Africa and sweep them across the Atlantic Ocean. Some of this fine particulate debris settles on boats that cruise the Caribbean Sea. Interestingly, scientists are just now discovering that there are grave ecological consequences associated with this massive influx of airborne dust. What follows are brief descriptions of some of these potential ecological effects.

■ Nutrient Input to Rainforests

Biologists have documented the fact that a regular airborne supply of essential nutrients, some thirteen million metric tons per year derived from African soils, drops onto the upper canopy of the Amazon rainforests. Rainwater washes the nutrients off the leaves onto the soil for uptake by the shallow tree roots. The fallout of these wind-blown fertilizers, in part, helps explain the incredible productivity and diversity of the plant communities of the Amazonian rainforests, despite the sterility of their leached soils.

■ Decline of Caribbean Coral Reefs

Dust plumes from northern Africa inject nutrients and other vital chemicals into the clear blue waters of the Caribbean Sea. This promotes the growth of seaweeds, which overgrow and suffocate the coral polyps. Also, 90 percent of the Caribbean-wide population of the sea urchin *Diadema antillarum* was recently killed by a disease allegedly transmitted by pathogens carried in dust plumes from North Africa. These sea urchins are voracious grazers, and they kept the coral clean of algae. Their collapse has allowed an uncontrolled growth of algae, which has overwhelmed and killed many corals of the Caribbean Sea. Marine biologists do not consider the widespread demise of Caribbean coral reefs and the unprecedented fallout in the area of dust from North Africa to be merely coincidental.

■ Red Tide Outbreaks

The wind-blown dust reaching the western Atlantic Ocean from Africa is iron rich and its fallout can change dramatically the chemistry of seawater. For example, the dissolved iron content of the waters of the Florida Keys has, on occasion, increased by as much as 300 percent following a dust-plume event. This infusion of iron has caused a large plankton bloom of bacteria, which in turn raised the nitrogen levels in the shallow water of the Florida Keys. The elevated nitrogen concentrations, marine biologists suspect, triggered a red tide (FIGURE B4–4c) that devastated local ecosystems, killing millions of fish and a hundred or so manatees, as well as making many people ill.

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Microbe Transport

An astonishing discovery is that a large variety of microorganisms can be wafted across the Atlantic Ocean in dust plumes. The soils of the Sahel region of Africa have a rich assortment of bacteria, viruses, fungi, and other microbial types. Some are pathogens and they may have infected and killed sea fans in the coral reefs of Florida and the Caribbean region. Recent kills of orange groves in Florida by infectious diseases likewise may be the result of the wind-blown transport of viruses from northern Africa.

Human Health Effects

The use of pesticides and herbicides and the burning of garbage with its load of plastics in northern Africa are releasing many toxins into the air. These chemicals get incorporated into the dust plumes that rain down in the southeastern United States and the Caribbean region. High levels of mercury, radioactive lead, and beryllium-7 have been measured in the fallout from the dust plumes emanating from northern Africa. The long-term health consequences for people of these chemical toxins are just beginning to be investigated. Even the dust itself is a health hazard, as indicated by the seventeen-fold increase in cases of asthma attacks in Barbados since 1973 when the air has been choked with African dust.

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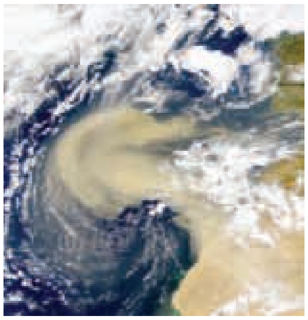
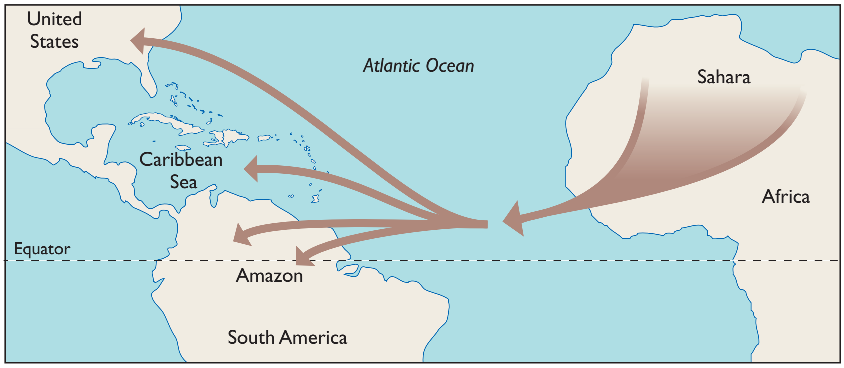

(a)



(b)


(c)

FIGURE B4–4 (a) This satellite image reveals the presence of dust plumes stretching from Africa far into the Atlantic Ocean. (b) The general flow paths of dust from the desertification of northern Africa. (c) The reddish hues in this photo reveal a red tide in Florida coastal waters. The fallout of wind-borne, iron-rich dust that originated in North Africa triggers some of these events.

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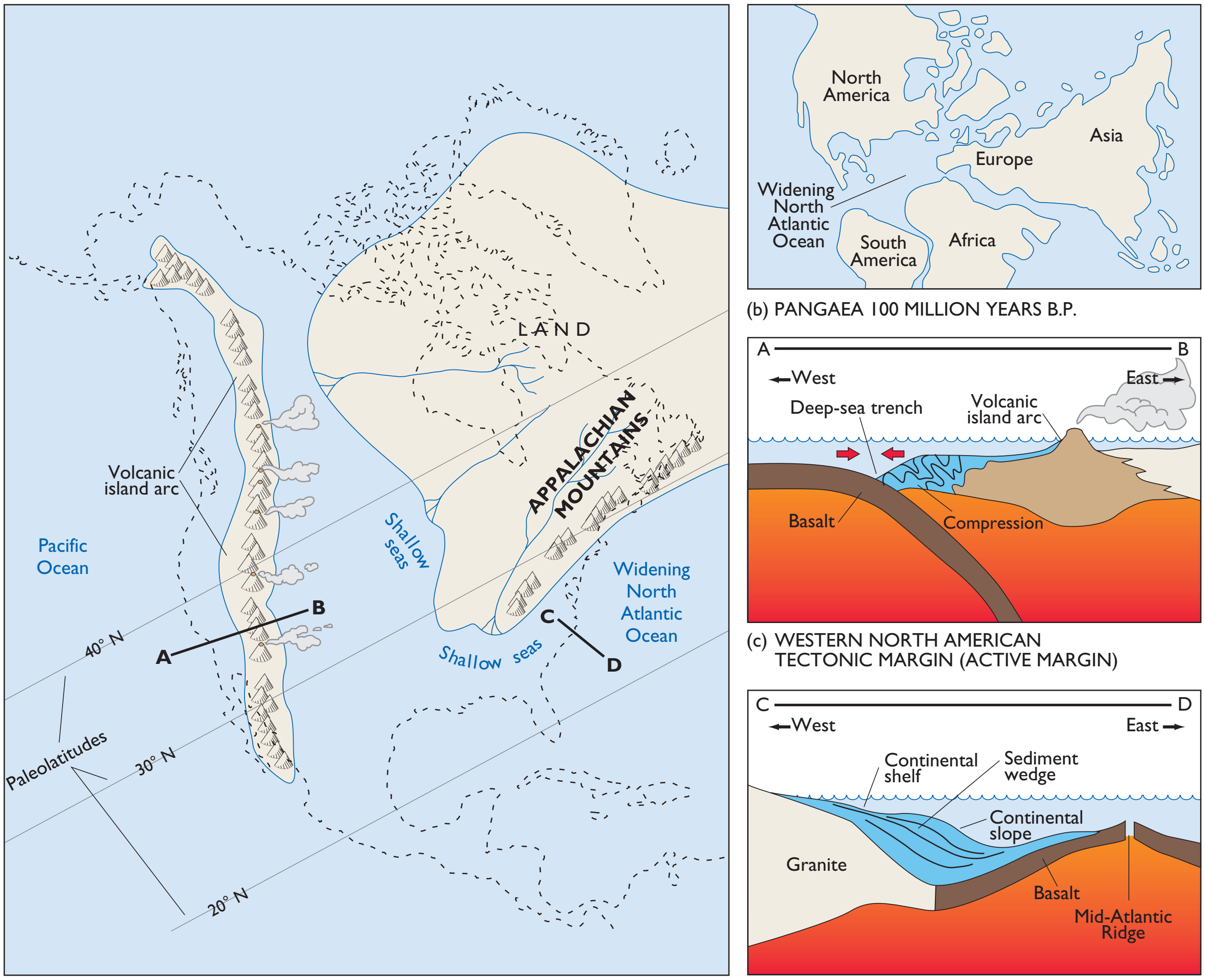
■ 10^8 TO 10^6 YEARS: The span of 100 million to 1 million years (10^8 to 10^6) is enormous and stretches far back into the geologic past to the age of the megacontinent Pangaea. The central and western parts of North America would have been unrecognizable to us today, as much of them were covered by warm, shallow seas, with a long chain of volcanic islands marking a complex subduction zone (FIGURE 4–5a). Recall that subduction zones are compressional boundaries (areas of "squeezing" together) where one plate is forced to dive beneath another plate. What is now the eastern edge of North America was a zone of tension (pulling apart), marked by fault-bounded rift valleys. At that time, North America had separated from Pangaea (FIGURE 4–5b), its eastern edge having just broken off from Africa, Europe, and Greenland. As Africa, Europe, and Greenland were moving eastward relative to North America because of seafloor spreading, a gulf was created between them and North America that widened with time into a young Atlantic Ocean. Thus, the North American continent had developed two distinct continental edges: a tectonic (active) Pacific-type of continental margin to the west, and a nontectonic



(a) NORTH AMERICAN PALEOGEOGRAPHY 100 MILLION YEARS B.P.

(b) PANGAEA 100 MILLION YEARS B.P.

(c) WESTERN NORTH AMERICAN TECTONIC MARGIN (ACTIVE MARGIN)

(d) EASTERN NORTH AMERICAN NONTECTONIC MARGIN (PASSIVE MARGIN)

FIGURE 4–5 **Paleogeography of North America.** (a) During the Cretaceous period (100 million years ago), the eastern edge of North America was undergoing rifting and sinking as the newly formed Atlantic Ocean was expanding. Most of central and western North America was covered by shallow seas, except for a prominent chain of volcanoes that marked a subduction zone where two plates were colliding. (Adapted from Dott, R. H., and Batten, R. L. *Evolution of the Earth*. McGraw-Hill, 1981.) (b) The relative position of landmasses during the Cretaceous period indicates that the juvenile North Atlantic Ocean was widening slowly during that time. (c) Western North America was a subduction boundary, with the Pacific plate sliding beneath North America. (d) Along the eastern edge of North America, sand and mud eroded off the Appalachian Mountains were deposited to form a thick and wide continental margin.

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(passive) Atlantic-type of continental margin to the east (FIGURES 4–5c and 4–5d).

Passive **Atlantic-type margins** are characterized by a long history of sedimentation. The seafloor at the edge of the continent sinks so gradually that the buildup of sediment keeps pace with the subsidence (sinking), and the shelf bottom remains shallow. Tectonic processes are important mostly during the earliest stage of evolution of passive continental margins. This is the time when continents are being ripped apart by plate tectonics (FIGURE 4–6a) and the crust at the new edge of the continent has been fractured, heated, and intruded by basalt rising along the newly formed spreading ridge. This early tectonic phase is comparatively short-lived, however, because the continents spread away from the active ridge. What follows is a long period of sedimentation. Terrigenous sediment eroded from the continent accumulates on the shelf, causing the continental margin to widen and its deposits to thicken (FIGURES 4–6b and 4–6c). With continued drifting away from the spreading ocean ridge, the crust and the underlying mantle at the passive edge of the continent

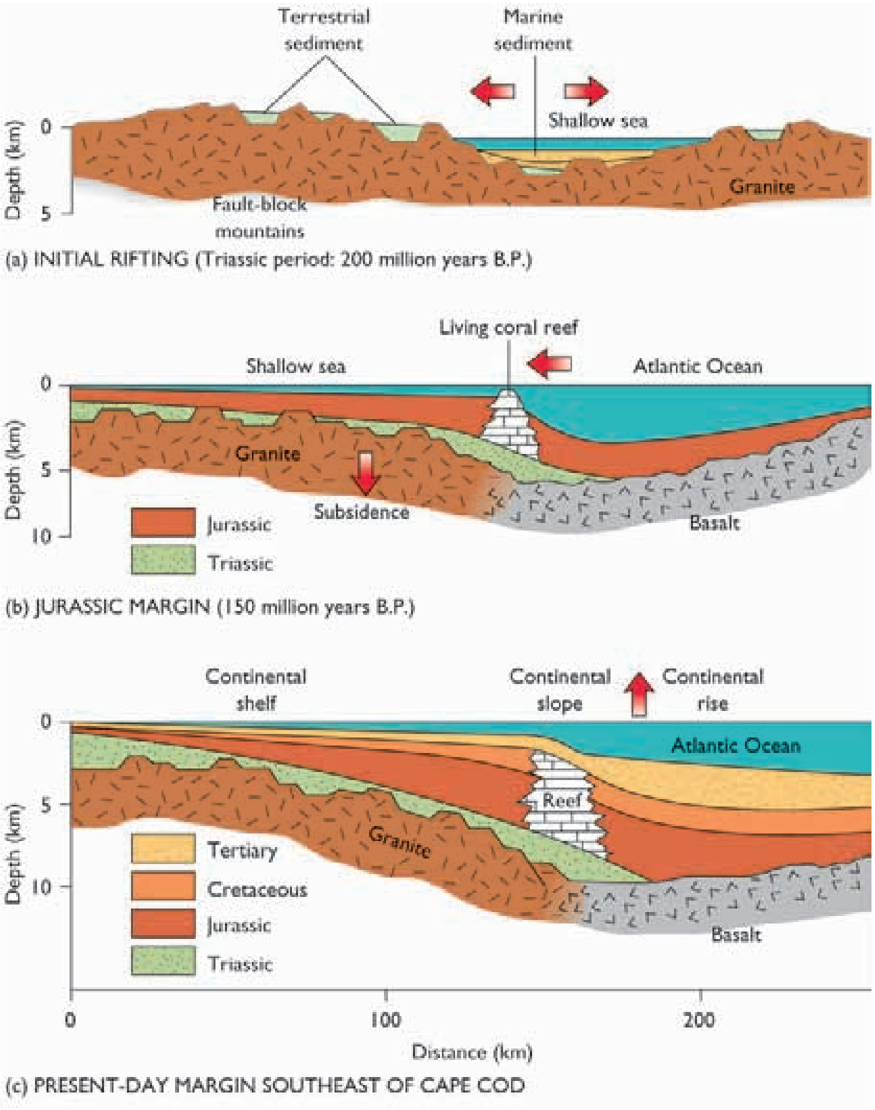

FIGURE 4–6 **Development of a passive Atlantic-type margin.** (a) The Atlantic continental margin of North America began when tension broke the crust into a series of fault-bounded ridges and basins. The rift basins were first filled in with river deposits and then with shallow marine deposits as seawater flooded the juvenile Atlantic basin. (b) Fifty million years after the initial rifting, subsidence of the granitic crust and deposition led to the development of a thick cover of marine sedimentary deposits. Far offshore, a thriving coral reef grew upward, keeping pace with crustal subsidence. (c) The present-day continental margin off New England shows a thick, broad sedimentary prism that completely buries the offshore reef, which died sometime during the Cretaceous period. (Adapted from Watkins, J. R., et al. *Geological and Geophysical Investigations of Continental*. American Association of Petroleum, 1979.)

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Climate Variability and Change

The Pleistocene Epoch consisted of a series of Ice Ages, each lasting about a hundred thousand years, when cold temperatures resulted in the buildup of huge ice masses on land. The Ice Ages were separated from one another by interglacial stages lasting between ten and twenty thousand years, during which ice sheets melted as the climate warmed worldwide. Because the most recent Ice Age ended some ten to fifteen thousand years ago, it is reasonable to suspect that the Earth may be poised at the brink of another Ice Age. Can science look into the future and test this possibility?

One approach at answering the question is to determine as accurately as possible the pattern of average global temperature over the past millennium and consider whether the trend can be justifiably extrapolated into the future. Accurate temperature measurements with thermometers go back only to the beginning of the twentieth century. Surface temperatures from earlier times are reconstructed from proxy evidence, such as tree-line positions, tree rings, the growth of coral reefs, and variations in lake and ocean sediments. As expected, the margin of error for establishing an average global temperature for the Earth increases the further back in time one goes.

Over the past thousand years, the Earth's surface temperature has varied significantly. There is a strong indication that an irregular cooling trend was under way from 1000 CE to about 1850 (FIGURE B4–5) although with wide geographic variability. This cooling trend ended the warmer "Medieval Little Optimum" that occurred between A.D. 900 and 1100. That was a time of sea exploration, and Norsemen colonized Greenland and Iceland because the Arctic pack ice lay far to the north and no longer impeded ship travel. Grain was grown in Iceland and Greenland at that time, and the local fisheries flourished. Beginning in the thirteenth and fourteenth centuries, temperatures began to drop globally, initiating the "Little Ice Age" (see Figure B4–5). Mountain glaciers advanced, the cover of sea ice in the North Atlantic expanded, snowfall increased, the snow cover lasted longer, and rivers froze during the winter as never before. The Norse settlements in Iceland and Greenland collapsed as grain harvests and fisheries failed, and expansive pack

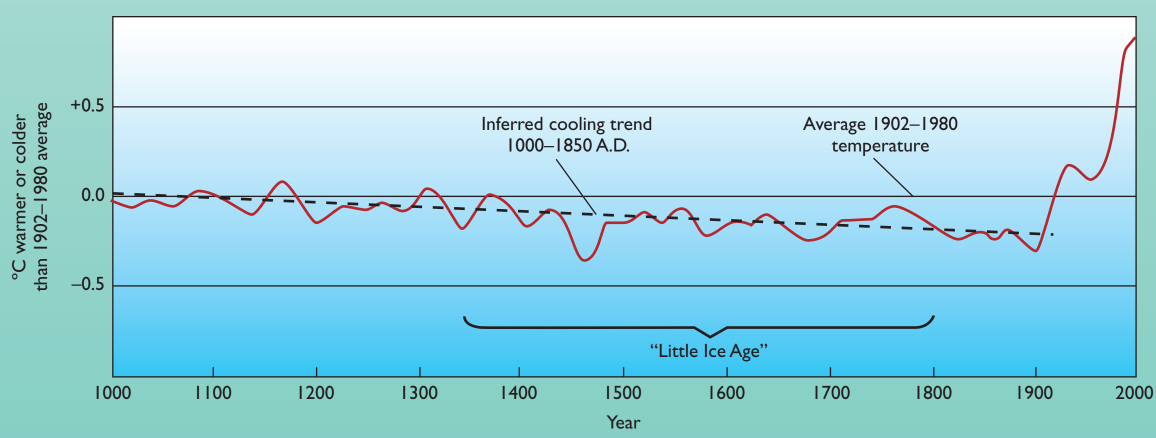

FIGURE B4–5 **Earth's surface temperature.** Since 1000 CE, the average global temperature has been dropping slowly. This trend was reversed in the early twentieth century as global surface temperatures began to rise sharply. Temperatures continue to rise, presumably due to atmospheric loading of heat-trapping gases such as carbon dioxide. [Adapted from Mann, M. E., et al. *Geophysical Research Letters* 26 (1999): 659–662.]

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ice in the North Atlantic Ocean prevented sea voyages between those colonies and Europe.

This nine hundred-year-long cooling suddenly ended in the early twentieth century; surface Earth temperatures have been rising sharply since then (see Figure B4–5). The general consensus among climatologists is that this temperature upswing is mostly a response to the buildup in the atmosphere of heat-trapping greenhouse gases such as carbon dioxide that have been released as a byproduct of the burning of fossil fuels. The carbon dioxide loading of the atmosphere by humans is expected to continue well into the twenty-first century. This suggests that the Earth's surface temperatures are going to continue to climb into the near future. At present, the global warming rate is estimated to be about 1°C every forty years, a rate that is much faster than it has probably been at any time in the past. So, rather than a prospect of chilly temperatures and the onset of an Ice Age, it appears as if the Earth will become warmer than it has been for thousands of years.

A large database of seawater temperature measurements indicates that 84% of the total heating of the Earth since the 1950s has occurred in the oceans. This heat gain has led to thermal expansion of the ocean, which accounts for at least 25% of the global sea level rise of the past 50 years. Based on computer models, many oceanographers believe that the warming of the oceans will enhance the stratification of the water column and modify deep-sea circulation patterns, including upwelling flows so crucial to marine biological productivity. Furthermore, many marine scientists attribute the thinning of polar ice shelves, the dramatic recent increase in the frequency and intensity of tropical storms, and the bleaching of coral reefs worldwide to global warming of the oceans. These phenomena are examined in detail in Chapter 16.

cool and contract and are weighed down by the sediment load. The edge of the continent sinks continually, providing ample room for the incoming loads of river sediment. The sand fraction of these river-supplied sediments is reworked by waves and currents and dispersed across the shelf. But the fine-mud load is kept in suspension and is moved seaward, where it may settle to the floor of the continental slope or bypass the continental margin entirely and eventually reach the abyssal depths of the deep-ocean floor. Study Figure 4–6 carefully; it demonstrates the development of the Atlantic continental margin of North America. The result of this long history of deposition is a broad, smooth continental shelf.

Now, let's contrast the Atlantic-type margin to the Pacific-type. Active **Pacific-type margins** not only receive sediment eroded from the nearby land, but also are affected by strong deformational forces. Subduction zones, you will recall, are boundaries where plates converge and are compressed. These strong tectonic forces squeeze the beds of sediment between the colliding plates, folding and faulting the sedimentary layers (FIGURE 4–7a). Also, the sedimentary layers and basalt are scraped off the top of the plate that is being forced downward by the upper plate. (This is roughly analogous to passing your hand across the top of a cake covered with frosting; as you do this, the icing piles up against your hand.) This tectonic activity creates an **accretionary prism** (FIGURE 4–7b)—a compressional zone situated between the deep-sea trench and the volcanic arc that widens with time as sediments are continually deformed and plastered to its seaward side by the subduction of plates. Along the nearby active volcanoes, sedimentary debris is derived from the erosion of the volcanic flows and deposited mostly underwater around the flanks of the volcanoes. Earthquakes abound here as well, and these trigger underwater landslides and slumps, which move large volumes of sedimentary debris to the deep-sea trench. Eventually, these materials get crushed against and are added to the accretionary prism. The result is a continental shelf that tends to be narrow and to have an irregular surface.

Carbonate Shelves

Because continents are drained by large river systems, a large supply of terrigenous debris is deposited onto continental shelves. Therefore, most shelves are covered by sand and mud composed of quartz and feldspar, the two principal minerals that comprise the granitic rock of the continents. Where the supply of river sediment is low and the waters of the shelf are warm, there can be a buildup of calcium carbonate ($CaCO_3$) material

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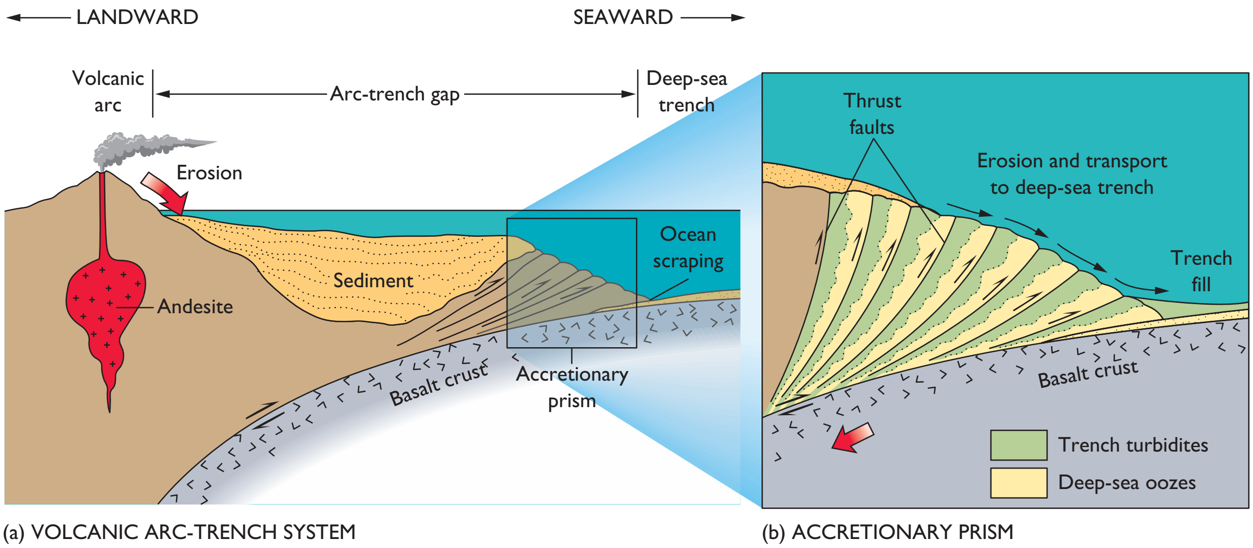

FIGURE 4–7 **Subduction tectonics and sedimentation.** (a) A subduction boundary includes three principal zones: the volcanic arc, the arc-trench gap (consisting of a sediment basin and an accretionary prism), and a deep-sea trench. Sediment is added to the arc-trench in two ways: (1) debris eroded from the volcanic arc slumps or is dumped by bottom currents into the ocean basin; or (2) deep-sea deposits and trench sediments are scraped off the descending plate and plastered onto the accretionary prism. (b) The accretionary prism widens with time as slices of trench turbidites and deep-sea oozes are added systematically to its base. (Adapted from Burke, C. A., and Drake, C. L. *The Geology of Continental Margins*. Springer-Verlag, 1974.)

composed of the shells of organisms. These shallow areas are called *carbonate shelves* or *platforms*. At present, carbonate sediment covers relatively few continental shelves of the world. Modern examples are located in tropical and subtropical oceans, such as those off southern Florida, the Bahamas, the Yucatán Peninsula of Mexico and nearby Central America, and northern Australia (FIGURE 4–8). The shelf water in these regions is clear, warm, shallow, and sunlit—conditions essential for the growth of abundant carbonate-secreting organisms. Moreover, if carbonate sediment is to accumulate, the input of terrigenous sand and mud must be minimal because muddy water not only interferes with the growth of reef organisms, but also dilutes the carbonate contribution by organisms to the bottom sediments. Therefore, carbonate shelves are located away from large rivers with their supply of terrigenous sediment. Carbonate deposits also accumulate around the shallow

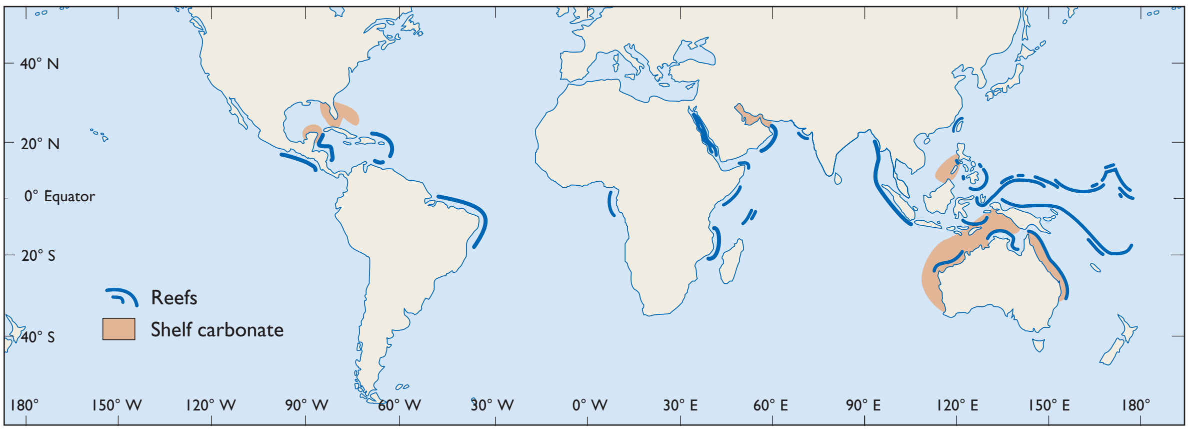

FIGURE 4–8 **Distribution of carbonate shelves.** Carbonate shelves are confined to tropical and subtropical settings, where the water is shallow, warm, and clear. Such sunlit environmental conditions foster the vast growth of carbonate-secreting organisms. The accumulation of carbonate sediment on the continental shelves occurs where the input of river sediment is negligible.

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edges of some islands, forming carbonate platforms; a fine example is the Bahama platform located east of southern Florida. The presence of thick and extensive sequences of carbonate rocks, such as **limestone** (cemented calcium-carbonate mud and sand), in the Rocky Mountains, the Alps, the Himalayas, the Andes, and other areas indicates that carbonate deposition was more widespread during certain geologic periods of the past than it is today. This shows how changeable environments are over long periods of time and how, in the geologic past, seas have repeatedly drowned the land and then retreated.

Deep-Sea Sedimentation

There are two main sources of sediment for the deep-ocean floor: (1) terrigenous mud and sand that bypass the shallow continental shelf and (2) the hard parts of surface-water microorganisms that settle to the deep-sea bottom. We will begin by examining a few details of these sediment sources and then talk about the way that sediment is dispersed along the floor of the deep sea. Then we will explore in some depth (pun intended!) how deep-sea sedimentation patterns are influenced by plate tectonics.

Sources of Sediment to the Deep Sea

Sediment that settles to the bottom of the deep sea is derived from either external or internal sources (FIGURE 4–9a). External sources are the terrigenous rocks of the land. Weathering, the chemical and mechanical disintegration of rock at or near the Earth's surface, breaks down the bedrock of the land into small particles—mainly sand and mud—that are transported to the oceans by rivers and winds. The major sources of terrigenous sediment in the oceans are rivers that drain large mountain belts, such as the Himalayas of Asia (FIGURE 4–9b). Here, steep, swift, and powerful rivers disgorge large quantities of sand and mud to the ocean. Internal sources of sediment furnish material that is produced largely by organisms and, to a lesser degree, by geochemical and biochemical precipitation of solids, such as ferromanganese nodules (hard pebbles enriched in metals). As a general rule, the proportion of deep-sea sediment derived from external sources (the terrigenous material) relative to that derived from internal sources (the biogenic material) decreases with distance offshore. In other words, the farther from the river supply, the greater tends to be the fraction of biogenic material in deep-sea deposits.

Next, let's examine the specific processes responsible for dispersing sediment to the floor of the deep sea. Once we understand these processes, we will be in a position to discuss patterns of sediment distribution in the open ocean far from the influence of land.

Sedimentation Processes in the Deep Sea

A simple classification of deep-sea deposits uses three broad categories based on the mode of sedimentation. **Bulk emplacement** is the means by which large quantities of sediment are transported to the deep-sea floor as a mass rather than as individual grains. The processes of bulk emplacement are induced by gravity: material resting high up on a slope moves downward and comes to rest on the floor of the deep sea. All types of sedimentary debris—terrigenous and biogenic, both fine- and coarse-grained—can be swept seaward and dispersed across the deep-ocean floor by bulk emplacement. In contrast, **pelagic sediment** is the fine-grained fallout of terrigenous and biogenic material that settles through the water column, particle by particle, much as snowflakes fall out of the sky and accumulate as a snow cover on land. **Hydrogenous sediment**, as you may recall, consists largely of biochemical precipitates that form in situ (in place); that is, they originate at the site of deposition by geochemical and biochemical reactions. Let's examine each type of sediment in more detail.

▪ BULK EMPLACEMENT: Terrigenous debris supplied by rivers enters the ocean along its edges, where most of it is deposited at the shoreline and the inner continental shelf. However, during low stands of sea level, rivers extend their channels seaward and drop sediment on the outer continental shelf and upper continental slope. See the featured box "Catastrophic Meltwater Scouring and Deposition," which examines how short-lived, extreme events on land can drastically affect the morphology of the seafloor. Buildup of such sediment can cause local instability and slope failure, which, under the influence of gravity, leads to downslope transport of sedimentary material as coherent slump masses, loose debris flows, or fluid mudflows. **Slumps** are sediment piles that slide downslope intact, with little internal deformation of the moving mass. In other words, bedding in the sediment pile is disturbed by folding, but is preserved when the whole mass of the deposit slides downhill as a sedimentary package. Examples of large slump masses are provided by seismic-reflection profiles (see the boxed feature, "Probing the Seafloor" in Chapter 2), which show large and small masses of crumpled sediment that must have slid downslope toward the deep sea (FIGURE 4–10a). Debris flows and mudflows are slurries, a mixture of

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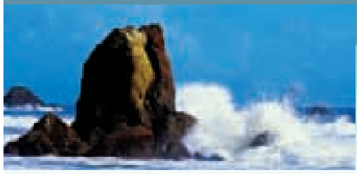

Geology

Catastrophic Meltwater Scouring and Deposition

From day to day, the processes that shape the land seem to operate independently of those that affect the ocean except along their coastal edges. After all, what do the abyssal depths of the deep sea have to do with the mountainous interiors of the land? Over long periods of time, however, oceans and continents exchange enormous volumes of sediment. Across geologic time scales, for example, tectonic forces crush deep-sea mudstones, limestones, and sandstones along compressional plate boundaries. These deformed strata get slowly uplifted into towering mountains, which then get worn down by glaciers and rivers that gradually sweep the eroded debris back to the ocean. This plate tectonic cycle is slow and requires hundreds of millions of years to complete. However, under certain conditions, extreme, short-lived events operating over a handful of weeks can affect markedly both the terrestrial and marine realms. Consider the catastrophic draining of a large glacial lake through the Hudson River Valley that separates New York from New England.

With the retreat of the last continental ice cap known as the Laurentide Ice Sheet, a series of glacial lakes—Glacial Lakes Vermont and Albany (FIGURE B4–6)—formed in the long, narrow Hudson River Valley situated between eastern New York state and western Vermont and Massachusetts. Farther north, glacial meltwater trapped between the Adirondack Mountains and the massive ice sheet to the north created Glacial Lake Iroquois, which was three times the size of Lake Ontario. About 13,350 years ago, an ice dam at the northeastern end of Lake Iroquois collapsed (Figure B4–6). In response, the lake level dropped some 120 meters (~400 feet), releasing instantly a large volume of water into the narrow confines of the Hudson River Valley. The raging torrent of floodwater scoured the valley of sediment, killing the flora and fauna, and transporting enormous boulders far past Manhattan, Brooklyn, and Staten Island onto the present-day continental shelf, reaching the head of the Hudson Submarine Canyon (Figure B4–6).

The effect of this short-lived event is still clearly apparent on the present-day seafloor south of New York and New Jersey. As the glacial floodwater breached a glacial moraine located at the Narrows in New York City, it moved tremendous quantities of debris onto the shelf and shaped them into a series of expansive sediment lobes. Large boulders, some with diameters greater than 2 m (~6.6 ft) and weighing over 2 tons, litter the seaward edges of these lobes, attesting to the power of the catastrophic flow needed to move them 50 to 100 km (~31 to 62 miles) across the shelf, which has a bottom gradient of less than one degree. Mastodon and mammoth teeth scattered about the flood debris of the lobes represent the remains of animals swept down the Hudson River Valley by the torrential flow. It's hard to imagine how quickly the ecosystems of the Hudson River Valley and the flat continental shelf, which was forested grassland at the time, were decimated by this extreme, short-lived event.

The story does not end there. The impact of this great flood may have affected global climate as well. The total freshwater discharge from draining Lake Iroquois is estimated to be at least 3.5×10^{12} m^3. This large pulse of glacial meltwater flooding down the Hudson River Valley and across the emergent sediment lobes of the shelf floated out to sea and likely affected ocean circulation. This sudden freshwater influx disrupted the normal sinking of salty surface water in the North Atlantic Ocean that, according to climate modelers, would be expected to trigger a short cooling event of global proportions by altering the exchange of heat between the atmosphere and ocean. There is both marine and terrestrial evidence for a well-documented, 400-year-long cold period known as the Intra-Allerod cold period that started 13,350 years ago, a time coincident with the catastrophic draining of Lake Iroquois. Some scientists believe that the timing between these two events is much more than coincidental. Scientists are currently testing this hypothesis by conducting additional field and theoretical work.

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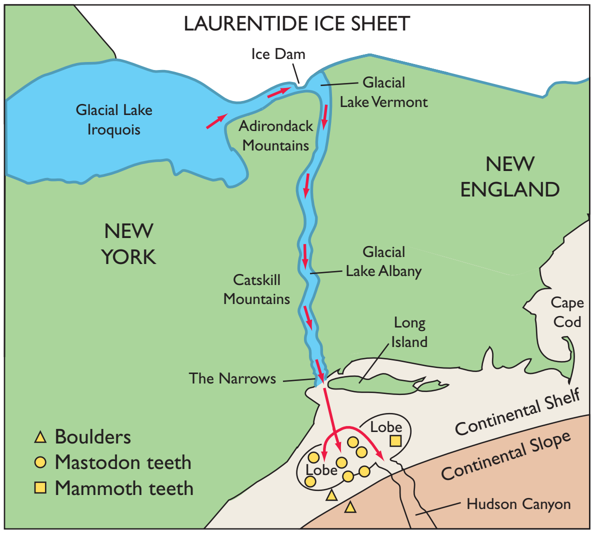

FIGURE B4–6 **Catastrophic drainage of Glacial Lake Iroquois.** The collapse of an ice dam around 13,350 years ago caused a raging torrent of water to scour the Hudson River Valley and dump sediment, large boulders, and mastadons on the present-day continental shelf south of Long Island. [Adapted from Donnelly, J. P., et al. *Geology* 33 (2005): 89–92.]

water and sediment that can sweep even large boulders downslope. Debris flows are mixtures of rock, sand, and mud; mudflows, of silt and clay. Unlike slumps, slurries destroy any previous bedding that may have existed in the deposits before they were disturbed.

Turbidity currents are important agents of transport to the deep sea. These powerful bottom currents are sediment-laden slurries that, under the influence of gravity, move rapidly downslope as turbulent underflows that push aside less dense water (FIGURE 4–10b). The slurry is created when sediment from the sea bottom is suspended and mixed with water. Imagine a large slump mass sliding downslope. The motion of the slump causes mud and sand to go into water suspension, and this slurry, too, begins to move downslope because of its high density. Once under way, a turbidity current becomes self-accelerating as it scours the sea bottom, placing more sediment into suspension and increasing its own density relative to the surrounding water even more, which causes its speed to increase. Where the sea bottom flattens and the turbidity currents slow, they deposit sediment rapidly and intermittently on the deep-sea bottom.

Large, deep submarine canyons, many with steep shoulders, are cut into the outer continental shelf and slope (see Figure 2–3a). For a long time, nobody could come up with a satisfactory explanation for their origin. The canyons appeared to have been eroded by rivers, but the parts of the canyons that cut into the outer shelf and continental slope are too deep in the ocean ever to have been subaerially exposed. It now appears that most submarine canyons have been excavated by a combination of sediment slumping and turbidity currents that have deepened a gully, or depression, on the sea bottom. Submarine canyons serve as chutes for funneling large quantities of terrigenous mud and sand from the shelf and slope into the deep sea by turbidity-current transport (FIGURE 4–10c).

Apparently, submarine canyons are quite active when sea level is low, as this allows rivers to form deltas at the heads of canyons. Much of this deltaic sediment is unstable and slumps, and this leads to the formation of turbidity currents, which then flow down the canyon axis. At the base of the continental slope, where the mouth of the submarine canyon opens onto the continental rise and the abyssal floor, the turbidity current is no longer channeled (confined by the shoulders of a canyon) and it decelerates because the bottom slope flattens out and spreads the flow. As the current slows down, grains are dropped out of suspension according to size—the largest first, the finest last—producing **graded bedding** (FIGURE 4–10d). The beds of sediment laid down by turbidity currents are called **turbidites**. These graded beds accumulate one on top of another, each representing a distinct turbidity-current flow. Turbidites abound at the mouth of submarine canyons, where they form thick sequences of cone-shaped deposits known as **deep-sea fans**. They are similar to

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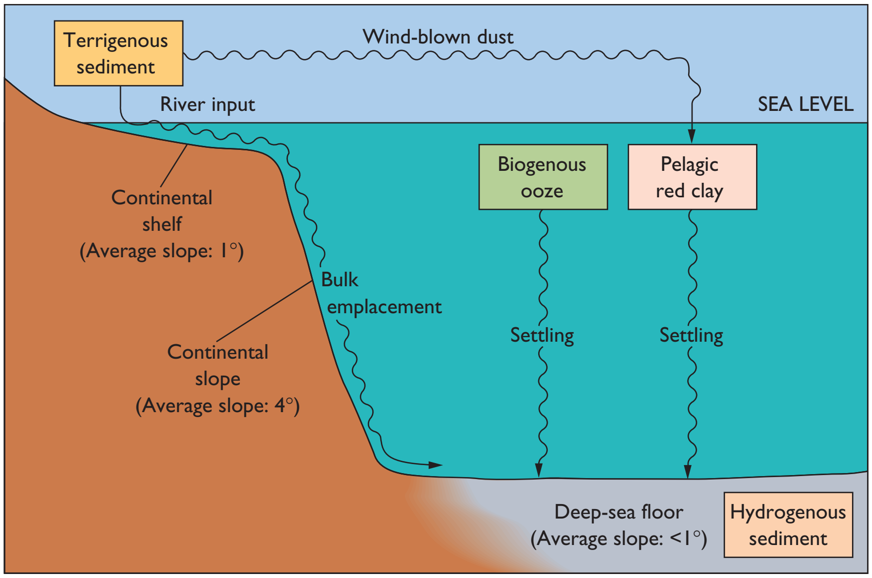
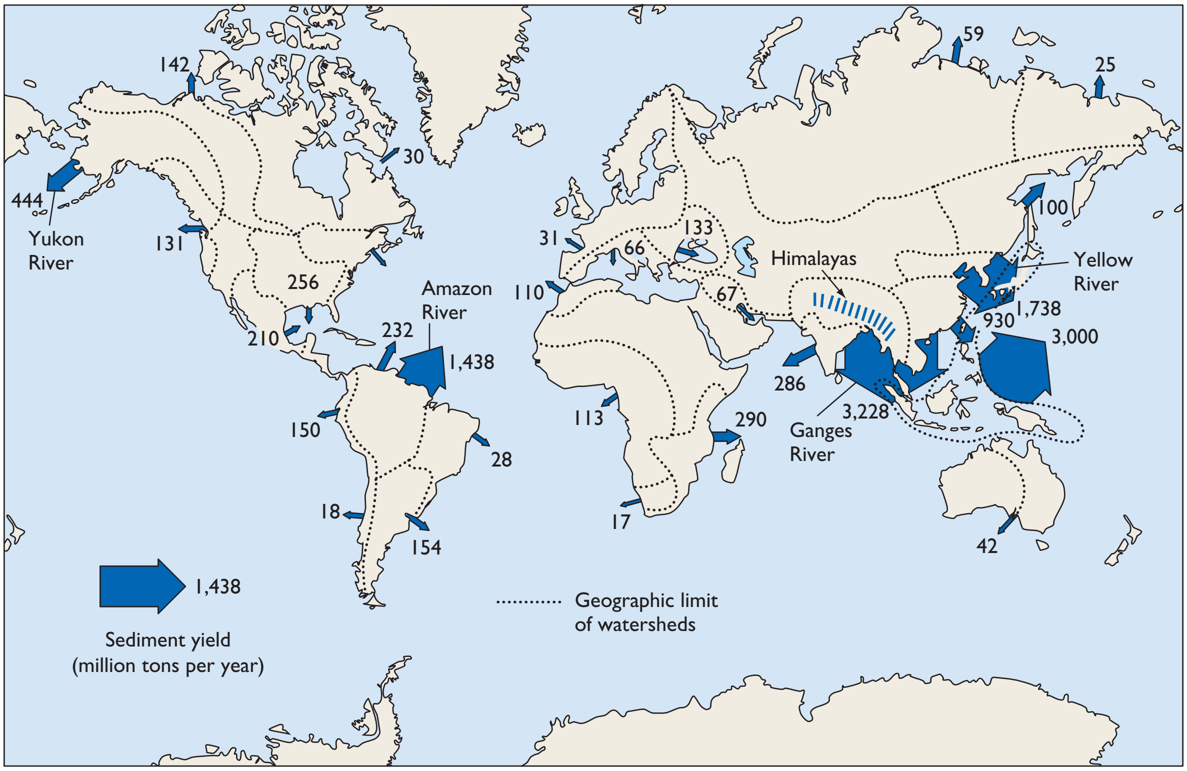


(a) SEDIMENTATION IN THE DEEP SEA



(b) RIVER INPUT OF SILT TO OCEANS

FIGURE 4–9 **Deep-sea sedimentation.** (a) Terrigenous sediment is derived from the breakdown of rocks on land and is supplied to the deep-sea by bulk emplacement. Pelagic sediment consists of oozes and red clay. The former is derived from biological production, the latter from erosion of rocks on land and transported as mud by rivers or by wind from land deserts. Hydrogenous sediments are chemical precipitates from seawater. (b) The influx of terrigenous debris to the ocean is quite variable. The vast bulk of it originates from the rivers that drain Asia. Another important source of river silt is the Amazon River of South America. [Adapted from Milliman, J. D., and Meade, R. H. *J Geol.* 91(1983): 1–21.]

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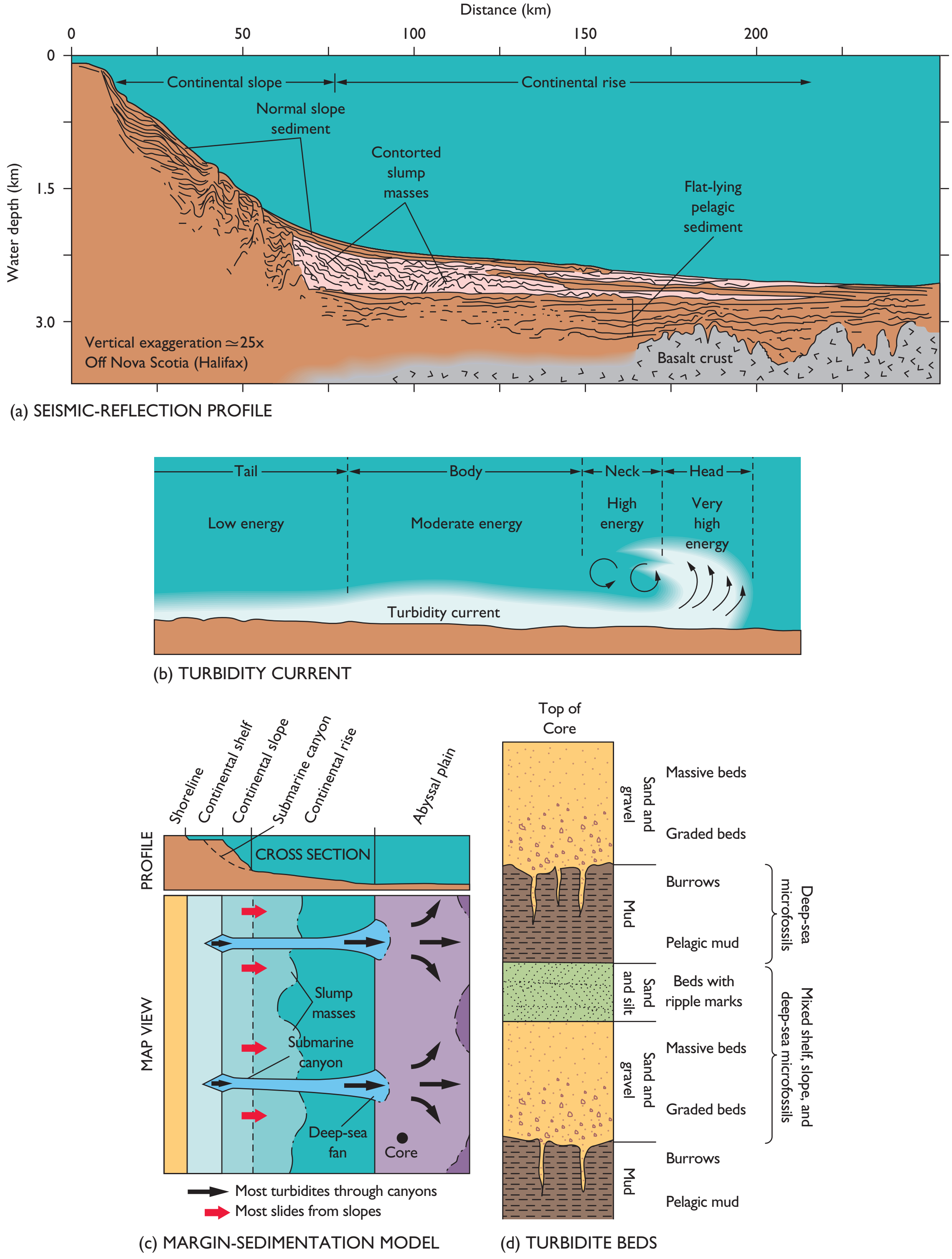

FIGURE 4–10 **Bulk emplacement of sediment to the deep sea.** (a) A seismic-reflection profile (see boxed feature, "Probing the Seafloor" in Chapter 2) taken along the Atlantic continental margin of Nova Scotia shows the lower slope and upper rise contain contorted, chaotic sediments. These deformed sediment masses are large slump blocks that were derived from somewhere upslope and slid downslope over flat-lying pelagic deposits. [Adapted from Emery, K. O., et. al. *AAPG Bulletin* 54 (1970): 44–108.] (b) A profile view of a turbidity current flowing down a slight incline because of its high density relative to the surrounding water. (c) Turbidity currents can carry coarse sand and gravel past the continental shelf and slope and down the floors of submarine canyons. These canyons serve as chutes that funnel sand onto the continental rise and the abyssal plain. [Adapted from Emery, K. O. *Oil and Gas Journal* 67 (1969): 231–243.] (d) Turbidity currents flow onto the deep-ocean floor, scouring the pelagic muds and depositing a sequence of sands and silts with characteristic graded bedding.

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FIGURE 4–11 **Desert alluvial fan.** Deep-sea fans resemble the alluvial fans that form at the mouths of steep canyons on land such as this one in the Anza-Borrego Desert State Park, California.

river deltas and to alluvial fans in desert environments (FIGURE 4–11). As the fan deposits grow, they may unite with other fans near them, creating the continental rise, and extend seaward, grading into the flat abyssal plain deposits (see Figure 4–10c).

The polar latitudes have their own style of bulk emplacement of terrigenous sediment to the deep sea. It is by the process of **ice rafting** (FIGURE 4–12a). Icebergs, which are large fragments of ice broken off from glaciers and ice sheets, may contain considerable amounts of sediment scraped off the land. Ocean currents transport these sediment-laden icebergs (rafts) away from land where, melting gradually, they drop their load of sediment into the deep sea. Ice rafting produces **glacial-marine sediments** (FIGURE 4–12b) that are characterized by poor sorting (a wide range of grain sizes, from boulders to clay particles) and a heterogeneous (nonuniform) composition of rock and mineral fragments.

▪ PELAGIC SEDIMENT: Most coarse terrigenous sediment—gravel and sand—supplied by rivers is deposited along the shoreline or the inner continental shelf. An exception previously discussed is the graded turbidite found on the deep-sea floor near the mouths of submarine canyons. The thick deposits of the continental rise consists of **hemipelagic sediment**, terrigenous mud, and silt that bypassed the continental shelf and slope and was reworked by bottom currents flowing parallel to the continental rise. Most of the deep-sea bottom, however, is blanketed by fine-grained mud composed of clay-sized and silt-sized particles that have settled slowly out of suspension in quiet, deep water far from the influence of land. These pelagic muds may be either inorganic or biogenic (organic) in origin; sometimes both types are mixed together. The former are mostly *red clays*, the latter are oozes. We will discuss both in detail.

The inorganic type of pelagic deposit is *red clay,* extremely fine-grained particles that typically have a brownish rather than reddish color, despite their name. In fact, many oceanographers simply call this **brown clay** or **pelagic clay**. These deposits are composed of various clay minerals, such as kaolinite and chlorite, and silt-sized and clay-sized grains of quartz and feldspar. Their color is the result of iron-bearing minerals that have been oxidized (producing, essentially, rust) by the oxygenated deep water. The precise origin of pelagic clay is uncertain. It appears to be derived from several sources, including the weathering of granitic and volcanic rocks, wind transport of dust from land, the fallout from space of extraterrestrial dust, and perhaps even the chemical precipitation of clays from seawater itself.

There is little doubt, however, that certain clay minerals in pelagic clay deposits originate by weathering of granitic rocks on land and are transported to the ocean in a variety of ways. Some are supplied by rivers, others are blown by the wind to the ocean, and still others are rafted by ice. To a large degree, climate in the source area controls the kind of clay minerals that form by weathering. For example, the warm, moist, acidic soils of the tropics favor the formation of kaolinite and the destruction of chlorite, two common species of clay. Given this relationship, the pelagic deposits of the low latitudes—the tropics and subtropics—are rich in kaolinite and poor in chlorite. This is indicated on the map of FIGURE 4–13 by a kaolinite/chlorite ratio that is very high. Yet, you may ask, what exactly is a kaolinite/chlorite ratio? Actually, it sounds technical, but it's quite a simple concept. A ratio of 1 indicates that the amount of the two clay minerals is identical because if you divide any number by itself you get a value of 1. A kaolinite/chlorite ratio greater than 1 signifies that there is more kaolinite than chlorite, so that you're dividing a number by a smaller value. For example, a kaolinite/chlorite ratio of 10 indicates that there is ten times more kaolinite than chlorite in the sediment sample. So if there are 10 grams of kaolinite, there must be 1 gram of chlorite. To test yourself,

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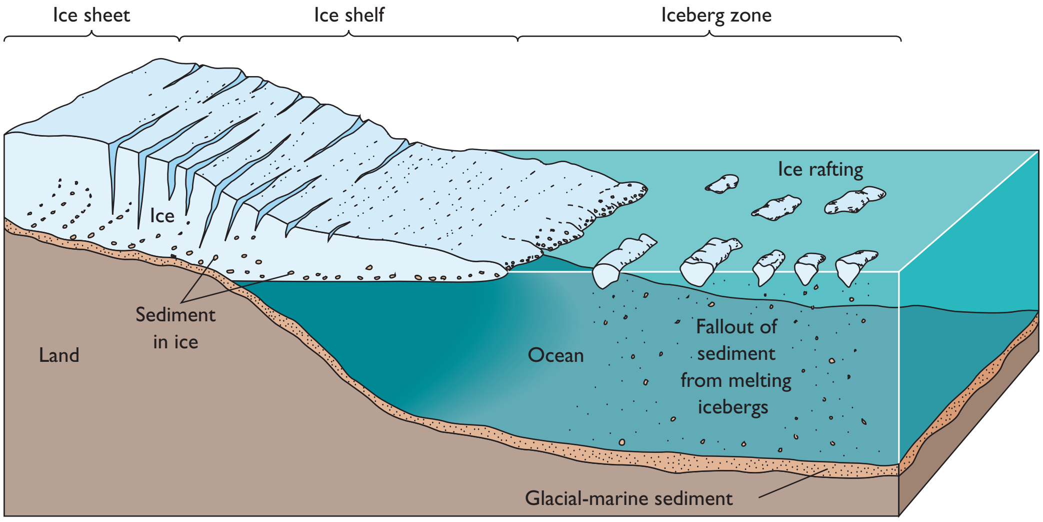
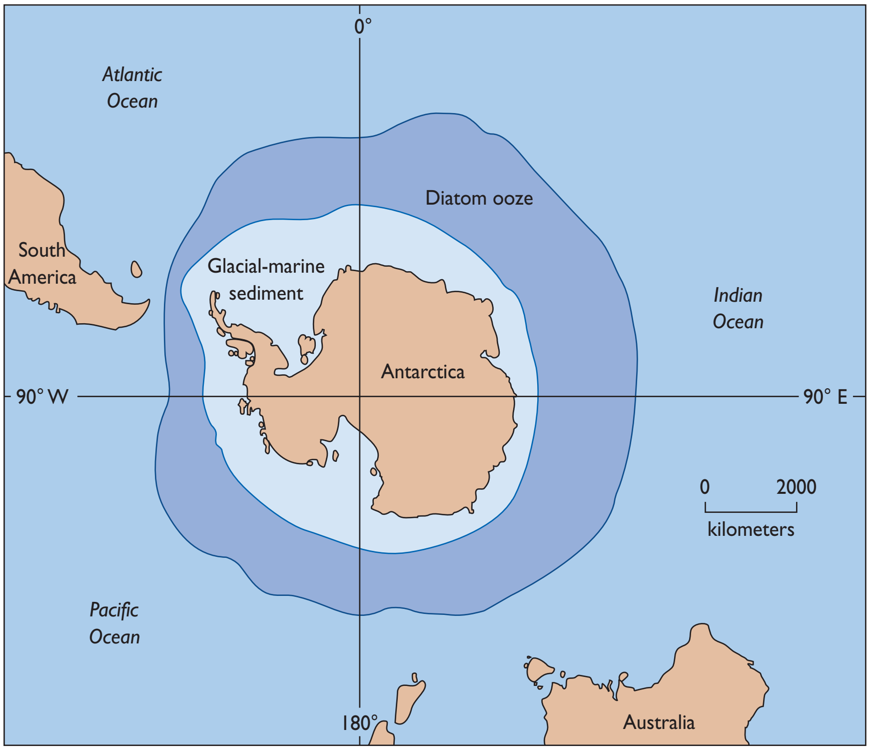


(a) ICE RAFTING



(b) DEEP-SEA DEPOSITS AROUND ANTARCTICA

FIGURE 4–12 **The formation of glacial-marine sediments.** (a) Glaciers extend into the ocean, where they form floating ice shelves that disintegrate into icebergs. The icebergs release their sediment load as they melt—a transport process termed ice rafting. (b) Ice rafting has created a 1,000-kilometer-broad band of glacial-marine sediment that surrounds Antarctica. (Adapted from Hays, J. D. *Progress in Oceanography*. Pergamon Press, 1967.)

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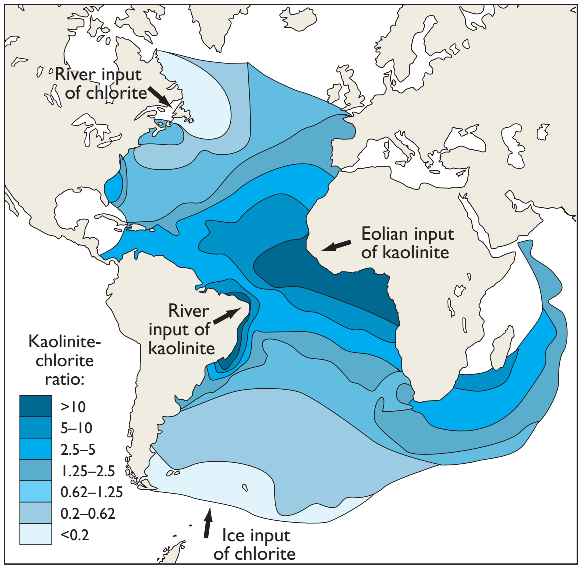

FIGURE 4–13 **Clays in deep-sea muds.** The kaolinite-chlorite ratio helps identify the source regions for much of these clay minerals because the quantity of a particular clay type decreases with distance from its source region. Kaolinite-chlorite ratios in pelagic deposits of the Atlantic Ocean reveal a distribution that is a function of latitude. Kaolinite abounds in the tropical latitudes; chlorite, in the polar latitudes. [Adapted from Biscaye, P. E. *Geol Soc Am Bull* 76 (1965): 810.]

answer this: What does a kaolinite/chlorite ratio of less than 1 denote? Now go to the map of Figure 4–13 and note that the highest kaolinite/chlorite ratios occur in the low latitudes, the lowest ratios in the high latitudes. This indicates that kaolinite predominates in the tropics and subtropics and chlorite in the temperate and subpolar latitudes. Also note that there are two "tongues" of pelagic sediment where the kaolinite/chlorite ratios are greater than 10. One of these is located off the Amazon River of South America and reflects the large quantity of kaolinite supplied to the deep sea by this very large river. The other is located offshore of the western Sahara Desert of Africa, where strong trade winds blow desert sand and dust as far westward as Barbados in the eastern Caribbean Sea—a distance of over 6,000 kilometers (~3,725 miles) (see boxed feature "Dust Storms")! Much of this wind-blown material is kaolinite, and it drops out of the air and settles to the bottom of the deep sea, helping to produce a band of deep-sea mud.

By definition, a **biogenous ooze** consists of 30 percent or more of the skeletal debris of microscopic organisms, most of which live in water far above the deep-sea floor, within a few hundred meters (a few hundred yards) of the ocean surface. The remaining 70 percent or less of the nonskeletal particles in oozes consists typically of inorganic mud particles. Biogenous deposits are divided into two major types according to their chemical composition: calcareous ($CaCO_3$) oozes and siliceous (SiO_2) oozes. **Calcareous oozes** are composed mainly of the tiny shells of **zooplankton** (floating single-celled animals), such as **foraminifera** and **pteropods**, and **phytoplankton** (floating single-celled plants), such as **coccolithophores** (FIGURE 4–14). Although abundances vary from area to area, all of these organisms are widely distributed in the surface water of the world's oceans. After the organisms die, their tiny hard parts dissolve or are incorporated into fecal pellets (the waste products of invertebrates) that settle slowly through the water column and eventually accumulate on the deep-sea floor. These shells may be dissolved, however, because cold bottom water tends to be slightly acidic (as will be explained in Chapter 5), and acid readily dissolves calcium carbonate. Oceanographers have defined the **carbonate compensation depth** (CCD) as the ocean level below which the preservation of $CaCO_3$ shells in surface sediments is negligible. The CCD depends on the rate of supply of carbonate and the acidity, temperature, and pressure of the water. Therefore, because the supply and dissolution rates of carbonate differ from place to place in the ocean, the exact depth of the CCD varies quite a bit; it tends to lie between 4 and 5 kilometers (~2.5 and 3.0 miles) below the sea surface. Rarely does carbonate ooze accumulate on ocean floor that is deeper than 5 kilometers (~3 miles). The CCD is an important chemical zone in the ocean that strongly controls the distribution of calcareous oozes.

Siliceous oozes consist of the remains of **diatoms** (floating, single-celled plants) and **radiolaria** (floating, single-celled animals) (see Figure 4–14). These organisms, which secrete hard parts made of silica (SiO_2), grow rapidly and are most abundant in water rich in nutrients. Silica is dissolved at very slow rates in seawater everywhere in the water column and tends to accumulate on the deep-sea bottom in areas where there is high biological productivity in the surface water. Two such regions of high biological production are sections of the polar and equatorial oceans; each produces a blanket of siliceous ooze on the underlying sea bottom. Typically, carbonate oozes accumulate on seafloor that is shallower than the CCD; silica oozes accumulate in the deep water below the CCD. Carbonate oozes are between three and nine times more abundant than are siliceous oozes in the world's ocean basins (see TABLE 4–2). This means that much of the ocean floor of the deep sea in each ocean basin lies well above the CCD; otherwise

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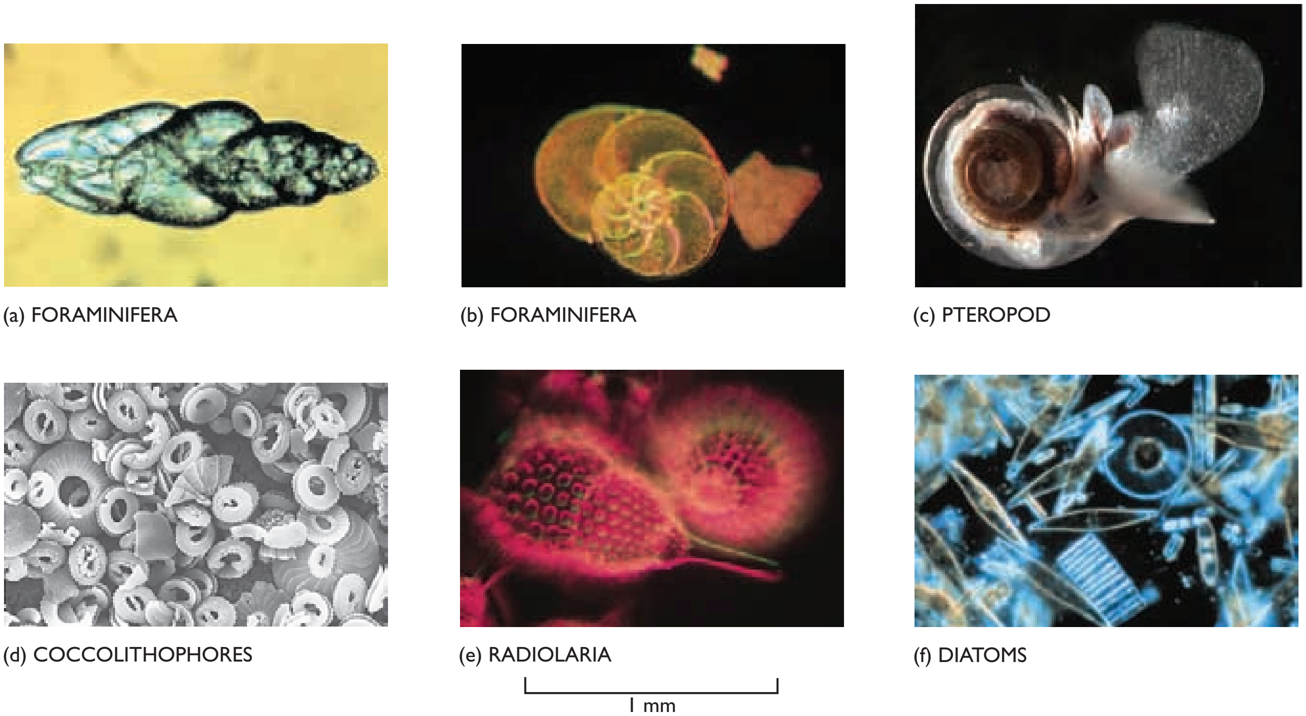

FIGURE 4–14 Common microfossils in biogenic oozes.

there would not be so much carbonate sediment on the floor of the deep sea. Table 4–2 shows the relative amounts of these various pelagic deposits. Note that globally 85 percent of pelagic sediment consists of clay (38 percent) and foraminiferal ooze (47 percent).

▪ HYDROGENOUS DEPOSITS: Authigenic deposits are chemical precipitates that form in place within an ocean basin. **Ferromanganese nodules** are one of the best-known examples of such material. These nodules (irregular to sphere-shaped masses) consist of concentric layers of various metal oxides (FIGURE 4–15a) that have precipitated around nuclei such as grains of sand, gravel, and even sharks' teeth. In addition to oxides of iron and manganese, which average about 20 to 30 percent by weight, ferromanganese nodules contain a variety of other metals, including nickel, copper, zinc, cobalt, and lead, making them a potentially valuable economic resource. It appears that the iron and manganese are derived from several sources, notably the chemical weathering of basalt, river-supplied compounds that are dissolved in water, and hot-water seepage (**hydrothermal**) from vents on the crest of spreading ocean ridges.

TABLE 4–2

DISTRIBUTION OF PELAGIC SEDIMENT

Type	Composition	Atlantic (%)	Pacific (%)	Indian (%)	Global (%)
Foraminiferal ooze	Carbonate	65	36	54	47
Pteropod ooze	Carbonate	2	0.1	—	0.5
Diatom ooze	Silica	7	10	20	12
Radiolarian ooze	Silica	—	5	0.5	3
Pelagic clay	Aluminum silicate	26	49	25	38

Source: Adapted from Berger, W. H. Biogenous deep sea sediments: production, preservation and interpretation. In *Chemical Oceanography*, vol. 5, Riley, J. P., and Chester, R., eds. (New York: Academic Press, 1976), 265–388; and Kennett, J. *Marine Geology* (Englewood Cliffs, N.J.: Prentice-Hall, 1982).

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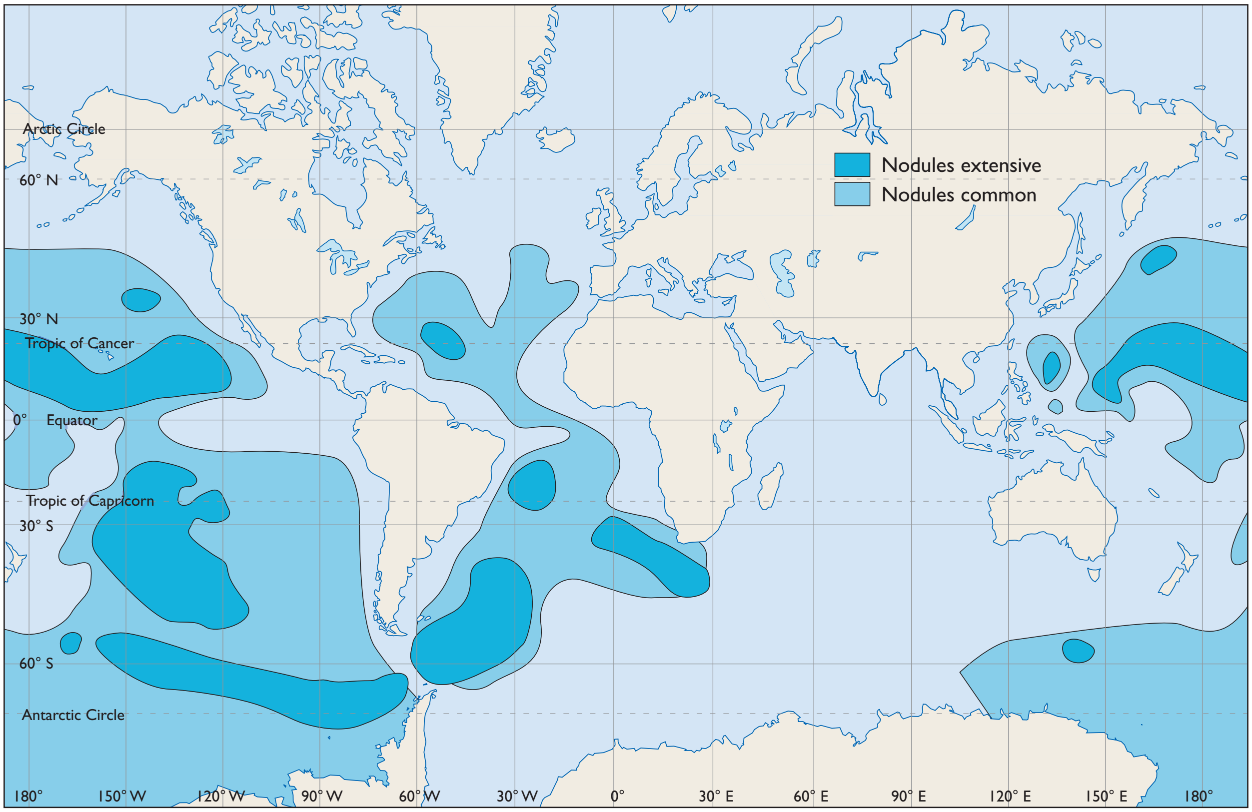
(a) PACIFIC NODULE



(b) GLOBAL DISTRIBUTION OF FERROMANGANESE NODULES

FIGURE 4–15 **Ferromanganese nodules.** (a) This photograph of a sectioned ferromanganese nodule reveals the concentric layering that typifies these authigenic deposits. (b) The global distribution of nodules shows that they are particularly abundant in the Pacific and South Atlantic Oceans. (Adapted from Cronan, D. S. *Marine Manganese Deposits*. Amsterdam: Elsevier, 1977.)

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The exact origin of these nodules is controversial. Some of the metals seem to have been concentrated in nodule layers by the activity of bacteria and foraminifera, which extract trace elements from the feces of microorganisms and transform them biochemically to a solid deposit on the nodule. Some geochemists, however, favor the theory of chemical precipitation directly from seawater or perhaps from hydrothermal submarine solutions. Whatever their origin, nodules clearly grow at the water-sediment interface (the contact zone between the water and the sediment). Also, they probably are rolled around, as indicated by their spherical shape. Because currents are usually weak in these areas, organisms burrowing in the sediment are believed to be the cause of nodules slowly shifting about over the sea bottom. Nodules grow very slowly, usually at rates of between 1 and 4 millimeters (~0.039 to 0.15 inches) per million years. Many reach the size of gravel and even baseballs. Although they can be found just about anywhere in the deep sea, they are particularly abundant in parts of the North and South Pacific and in the South Atlantic (FIGURE 4–15b).

Phosphorite, another mineral deposit of possible economic importance, is composed of up to 30 percent P_2O_5 by weight. Unlike the ferromanganese nodules of the deep-ocean floor, phosphorite nodules generally are restricted to the continental shelf and upper continental slope, where cold water rich in nutrients typically upwells (moves upward) to the surface. This upwelling water creates considerable biological productivity, and large quantities of matter rich in organic phosphate settle to the ocean floor. Once buried in sediment, the unoxidized organic detritus is eventually transformed into phosphorite, which can grow as nodules at a rate of 1 to 10 millimeters (~0.04 to 0.39 inches) per thousand years.

Other important chemical precipitates in the deep sea are the metal sulfide deposits discovered around hydrothermal vents of actively spreading ocean-ridge crests. The hot water that escapes from these volcanic vents is loaded with dissolved metals that are precipitated on the adjacent ocean floor as cooling and chemical adjustment take place.

Global Distribution of Deep-Sea Sediments

We will now survey the distribution of sediments across the seafloor and examine their variation with depth below the sea bottom. Once sediment is buried, the grains are compacted, cemented, and slowly transformed into sedimentary rock. Sand becomes **sandstone**; mud becomes either **shale**, if composed of clay minerals, or **limestone**, if composed of carbonate ooze.

Surface Deposits

FIGURE 4–16a is a map of the sedimentary deposits that blanket the deep-sea floor. The clear distribution patterns that are evident reflect the source of these various materials. The continents are the principal suppliers of terrigenous debris. The bulk of this material that is supplied to the ocean by rivers is deposited on the continental margins. A small fraction of the terrigenous sediment bypasses the margins and is dispersed into the deep sea by downslope slumping, debris and mud flows, and turbidity currents. These processes have provided terrigenous sediment to the abyssal plains of the North Atlantic Ocean and of the Indian Ocean on either side of India (see Figure 4–16a). Few such deposits are evident in the Pacific Ocean because this basin is surrounded by deep-sea trenches—deep elongated basins that trap any terrigenous sediment, such as turbidites, that manages to bypass the continental shelf. In the polar seas, ice rafting introduces terrigenous sediment to the deep sea. These glacial-marine deposits derived from melting icebergs are evident north of Iceland and in a broad band that circles the Antarctic continent (see Figures 4–12b and 4–16a).

Toward the center of all the ocean basins, far away from continental inputs of terrigenous debris, the seafloor is blanketed by pelagic deposits. Areas that have high biological productivity support large populations of planktonic organisms that contribute large quantities of siliceous and calcareous shells to the deep-sea bottom. The fertile polar seas favor the formation of diatom oozes, evident especially in the northern Pacific Ocean and off Antarctica seaward of the band of glacial-marine deposits (see Figure 4–12b). The high biological productivity of the equator produces a band of siliceous ooze in this region as well.

Calcareous oozes accumulate in water depths above the CCD and cover bathymetric highs, such as the crest and flanks of the spreading ocean ridges, the tops of seamounts, and the shallow, broad plateaus of the southwestern Pacific Ocean (see Figure 4–16a). Pelagic clay, which accumulates very slowly in the deep sea, forms in quiet environments far from other sources of sediment, such as terrigenous debris and biogenic oozes, and in deep water where calcareous particles are dissolved. Therefore, clay deposits are usually found in the deepest parts of the ocean basins, below the CCD and away from continents and areas of high surface productivity. Actually, clay particles are deposited in most areas of the deepest sea, but at such slow rates that they are diluted by the abundance of other sedimentary components. Clay deposits are most extensive in the Pacific Ocean. This is an ocean basin that is geologi-

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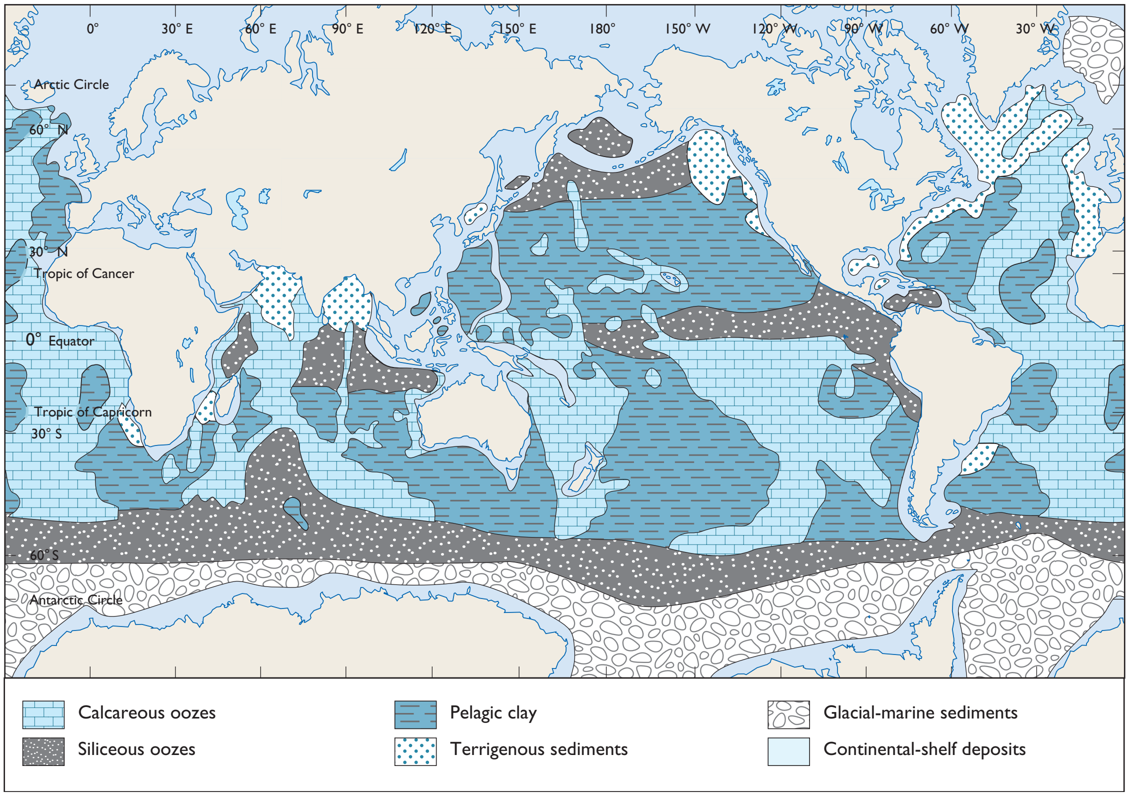
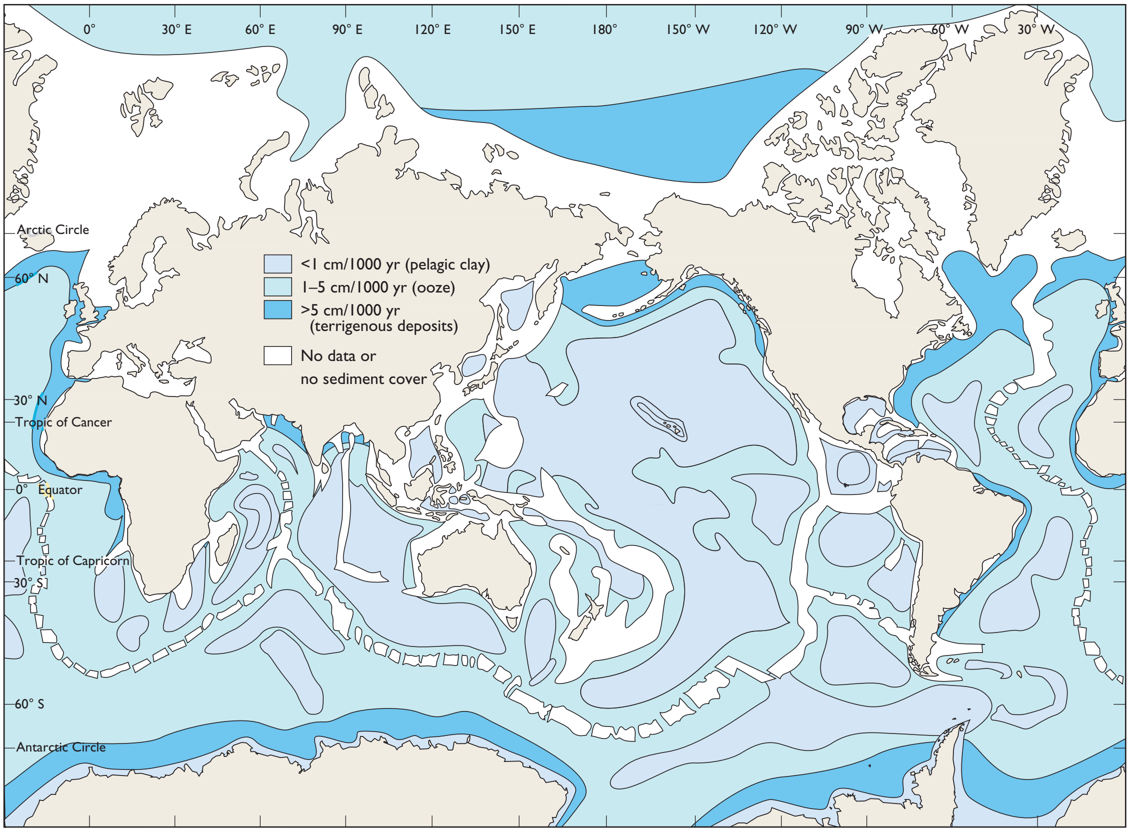


(a) DEEP-SEA SEDIMENT DISTRIBUTION



(b) SEDIMENTATION RATES

FIGURE 4–16 **Global deep-sea deposits.** (a) Deep-sea deposits vary worldwide with distance from land and water depth. (Adapted from Davies, T. A., et al. *Chemical Oceanography*. Academic Press, 1976.) (b) A comparison map shows that terrigenous deposits have the highest sedimentation rates and red clays have the lowest. Oozes have sedimentation rates that lie between these two extremes. (Adapted from Heezen, B. C., and Hollister, C. D. *The Face of the Deep*. Oxford University Press, 1971.)

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cally old and deep (below the CCD) and has extensive areas of low biological productivity. Thus, siliceous and carbonate oozes are uncommon and do not dilute the red or brown pelagic clays that accumulate in much of the Pacific.

Rates of sedimentation in the ocean basins vary with the composition of the sedimentary material (FIGURE 4–16b). The highest rates in the deep sea exceed 5 centimeters (~2 inches) per 1,000 years and are associated with the thick terrigenous deposits of the continental margins. Biogenic oozes accumulate in the deep sea at rates of 1 to 5 centimeters (~0.4 to 1.8 inches) per 1,000 years. The slowest depositional rates, which are less than 1 centimeter (~0.4 inch) per 1,000 years are associated with the pelagic clays that lie in the remotest parts of the ocean basins at depths well below the CCD.

Deep-Sea Stratigraphy

Seafloor spreading means that new ocean floor is created at the ocean ridge crests and then spreads away. During its long journey away from the axis of the Mid-Atlantic Ridge, the ocean floor of the North Atlantic basin has been receiving a steady supply of sediment at a rate of between 1 and 5 centimeters (~0.4 and 2 inches) per 1,000 years. Near the ridge crest, the basalt seafloor is bare, having just been formed. With distance down the flank of the ridge, the sediment cover thickens. In fact, there is a good correlation between the age of the basalt seafloor and the thickness of the sedimentary cover (FIGURE 4–17). At the very top of the ridge—far from sources of terrigenous sediment and well above the CCD—foraminiferal (foram for short) oozes accumulate and cover the basalt rock. With time, these oozes are buried and solidify into limestone. The biogenous deposits thicken with distance from the ridge axis because the underlying seafloor becomes older away from the ridge crest and has been subjected to a longer history of pelagic sedimentation. As the plate spreads and it cools and subsides below the CCD, the foram shells that reach the seafloor dissolve, and only pelagic mud accumulates, burying and protecting the older limestones. Consequently, a two-layered sequence develops, consisting of a top layer of mud and an older bottom layer of limestone (see Figure 4–17). Oceanographers refer to this as a simple layer-cake **stratigraphy** (stratified or bedded rocks).

FIGURE 4–18 is an interpretation of the deep-sea stratigraphy of the central Pacific Ocean. Notice that it is more complex than the simple two-layered stratigraphy of the Atlantic Ocean. This is because of the orientation of the East Pacific Rise (the active spreading ridge of the Pacific Ocean). Unlike the Atlantic plates, which are spreading east-west, parallel to lines of latitude, the Pacific seafloor is spreading west-northwest and east-southeast. This means that the seafloor of the Pacific is crossing climatic belts as it spreads, producing a more complex stratigraphy as sediment accumulates. Note that the two deepest sedimentary units that lie just above the basalt crust are identical to the deposits of the Atlantic basin (see FIGURE 4–18b). A thick layer of basal limestone that accumulated on the crest and upper flank of the East Pacific Rise above the CCD is overlain by shale (mud compacted into rock) produced by the settling of mud particles on the lower flank of the ridge below the CCD. Because of seafloor spreading,

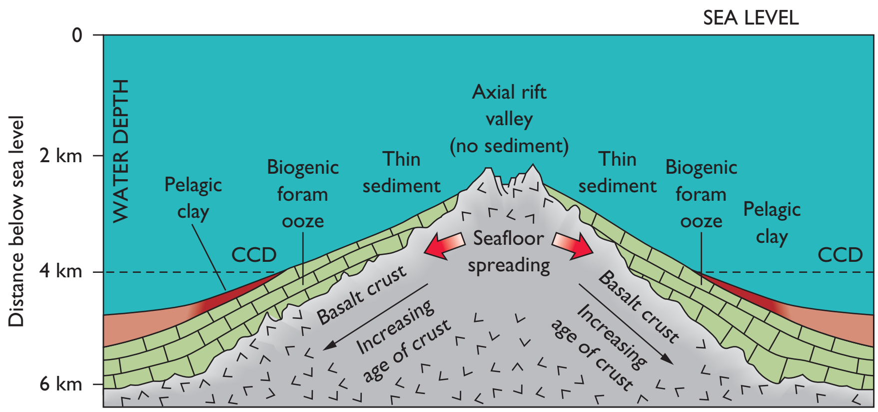

FIGURE 4–17 **Stratigraphy of the Atlantic Basin.** The deposits that cover the basalt crust form a "layer cake" stratigraphy: a thick layer of basal ooze, formed above the CCD (carbonate compensation depth), is overlain by a thinner mud layer that accumulates below the CCD. Note that there is no sediment at the crest of the midocean ridge and that the sediment cover thins away from the crest. This pattern is the result of seafloor spreading. As the basalt crust spreads away from the ridge crest, it acquires an ever thicker cover of sediment.

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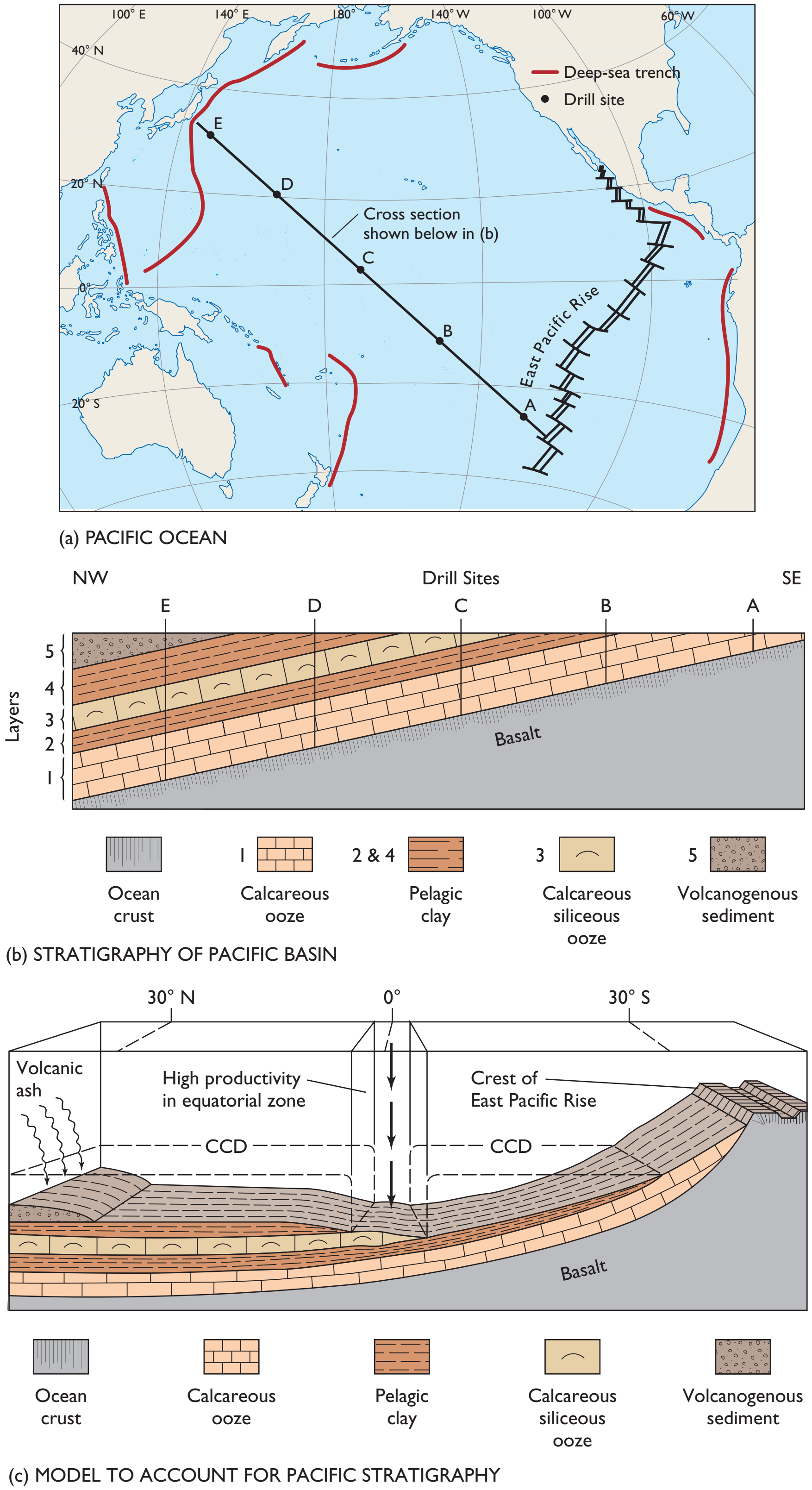

FIGURE 4–18 **Stratigraphy of the Pacific Basin.** (a) This generalized map of the Pacific Basin shows the location of the stratigraphic cross section in part b. (b) Because the floor of the Pacific Ocean is spreading to the northwest across lines of latitude, the stratigraphy is more complex than the two-layered stratigraphy of the Atlantic (see Figure 4–17). The letters are keyed to the drill sites shown in part a. [Adapted from Heezen, B. C., et al. *Nature* 241 (1973): 25–32.] (c) A basal layer of limestone forms on the flank of the ridge above the CCD. This limestone is topped by pelagic mud when the ocean floor sinks below the CCD, creating the simple two-layered stratigraphy evident in the Atlantic Ocean. As the ocean floor moves across the fertile equatorial belt, a second layer of limestone, rich in calcium-carbonate and siliceous fossils, accumulates and then becomes covered by a second deposit of pelagic mud north of the equator. Finally, a fifth layer of volcanogenous debris accumulates as the spreading seafloor approaches the volcanic arc of a subduction zone. [Adapted from Heezen, B. C., and MacGregor, I. D. *Sci Am* 229 (1973):102–112.]

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the Pacific plate crosses the equator, a fertile area that produces a great deal of biogenic ooze (FIGURE 4–18c). As a consequence of this passage across the equatorial zone, another limestone layer was produced that buried the shale. The very high biological productivity of equatorial waters causes the CCD to be lowered to the seafloor, allowing carbonate shells to accumulate at water depths below 4 kilometers (~2.5 miles). Once the Pacific sea bottom passed through the fertile equatorial zone and the CCD returned to a normal depth of 4 kilometers, a mud layer was deposited, forming a fourth stratigraphic unit (see Figures 4–18b and 4–18c). Finally, as the seafloor approaches the trench of a subduction zone, it may receive a supply of volcanic debris from the island arc, creating a localized fifth layer of volcanogenous sediment that overtops the older sedimentary layers.

Tectonics at subduction zones also influence the composition and stratigraphy of deep-sea deposits. One such example is described in the boxed feature, "The Drying Up of the Mediterranean Sea."

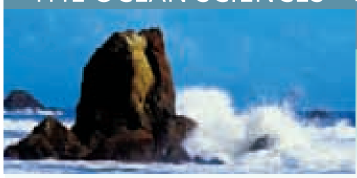
4-3 Future Discoveries

Sediment accumulates on the sea bottom and its layers represent a remarkable historical record of the geologic past. Exciting studies of sediment cores taken from the sea bottom all over the world are under way to document variations in climate and fluctuations in sea level extending back several hundred million years into the geologic past. By understanding the factors that possibly induced climate changes during earlier times, scientists will be in a better position to predict future climate and to anticipate the effects on life forms and processes that such worldwide changes are likely to produce.

Important questions that can be addressed by examining the sedimentary deposits of the ocean include: How did the ocean-atmosphere system respond to global warming events of the geologic past? How sensitive was the marine biota to extreme climatic changes in the past? After global catastrophic events, how long did it take marine plants and animals to rebound, and why did some groups recover more quickly than others?

Other marine geologists are testing the theory that some mass extinctions in the geologic past, such as the disappearance of dinosaurs 65 million years ago, may have been the result of giant meteorites striking the Earth. Scientists are searching the sedimentary record of the sea bottom for evidence of meteorite collisions with the Earth and looking for the impact craters themselves. For example, a meteorite impact presumably caused the extinction of the dinosaurs 65 million years ago, which marked the end of the Cretaceous Period and the onset of the Tertiary Period. Subsequently, an impact crater of the right age was found offshore of the Yucatan Peninsula, Mexico, in the Gulf of Mexico. Drilling off Florida by the Ocean Drilling Program sampled the Cretaceous-Tertiary boundary and found impact-shocked quartz, tektites, a high iridium content derived from a meteorite, and even a chunk of reef rock blasted into the air from the Yucatan. Clearly the dinosaurs expired in part because of this catastrophic event. These important findings will undoubtedly change our views of the Earth's history and the evolution of its marine and terrestrial biota.

THE OCEAN SCIENCES



Geology

The Drying Up of the Mediterranean Sea

The Romans coined the term "Mediterranean" because, to them, the region with its landlocked sea represented the very middle of the earth. The Mediterranean Sea is surrounded by three continents (Africa, Asia, and Europe) and is connected to the Atlantic Ocean by a narrow seaway, the Strait of Gibraltar (FIGURE B4–7). The northern edge of the Mediterranean Sea is dotted with many volcanoes and underlain by rock that has been squeezed tightly. These features, together with the abundant earthquakes in the region, indicate that this is a subduction zone along which two lithospheric plates are colliding and crushing each other. Africa, as it has been doing for hundreds of millions of years, is drifting slowly northward and encroaching on Europe and, in the process, is folding and uplifting marine sedimentary beds to form the Alps.

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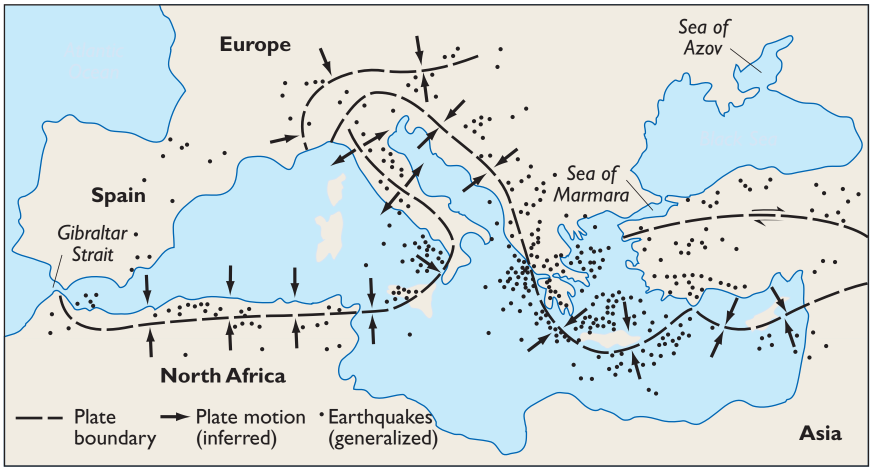

FIGURE B4–7 **Plate tectonics in the Mediterranean region.** The Mediterranean area is a tectonically active subduction zone where plates are colliding. The Mediterranean and Black Seas are the remnants of a much larger ocean that is slowly contracting as Africa drifts northward into Europe.

In 1972, deep-sea drilling in the western Mediterranean Sea by the *Glomar Challenger* uncovered unusual sedimentary deposits of Miocene age (5 to 25 million years old). Specifically, these deposits were anhydrite and stromatolites. **Anhydrite** (calcium sulfate) forms in hot, arid desert regions where saline water in the ground, but close to the surface, evaporates and causes calcium sulfate to precipitate out of solution. How can evaporites that form on dry desert land be part of the thick sedimentary deposits that lie on the bottom of the Mediterranean Sea? This is a very intriguing question indeed.

Stromatolites are also unrelated to deep-sea environments. These are crinkled laminations of carbonate mud. Today, stromatolite deposits are forming in the broad, intertidal mud flats of the Bahama Islands off Florida and in certain salty bays in Australia. There, dense mats of algae grow on the bottom and trap mud. The distinctive, crinkled structure of the mud laminations results from the irregular surface of the algal mat. The important point to remember is that stromatolites indicate shallow-water—not deep-water—deposition.

To Kenneth Hsü, the chief scientist aboard the *Glomar Challenger*, the conclusion was inescapable. Miocene deposits now lying on the deep-sea bottom of the Mediterranean Sea must have accumulated in desert environments (anhydrite) and in salty, shallow seas (stromatolites). This seemed preposterous at the time because the sedimentary units that immediately overlie and underlie these Miocene deposits are unmistakably deep-sea carbonate oozes. The sequence implies that the floor of the Mediterranean Sea must have been shoved up thousands of meters during the Miocene epoch, drained of its seawater and, shortly thereafter, must have dropped several thousand meters to abyssal depths—as if the basin were a veritable tectonic yo-yo (FIGURE B4–8a). Yet there is another way to interpret this story. The crust may not have moved up or down at all. Rather, the Mediterranean Sea may have dried out during the Miocene and then refilled quickly with seawater. In this case, the only fact for researchers to explain is how the Mediterranean Sea could have been drained of its water (FIGURE B4–8b)! This question is less difficult to answer than one might expect.

Only two conditions are necessary for the Mediterranean Sea to dry out entirely. First, the inflow of seawater from the Atlantic Ocean through the Strait of Gibraltar must be blocked. It turns out that this is easy to do, because the strait is a rock sill with its top only 100 meters (~330 feet) deep. Second, the climate needs to be arid to evaporate all the water trapped in the basin. The rock record indicates that the climate was hot and dry in the Mediterranean region during the Miocene epoch. Hsü calculated that it would have taken about a thousand years to evaporate all of the water trapped in the basin, once the sill had been raised and cut off the inflow of Atlantic seawater. This means that a sea would have been converted into a desert and back into a sea during a single millennium!

When the western Mediterranean had dried out, the seafloor would have been converted into a desert terrain with isolated salty lakes (FIGURE B4–9a). The

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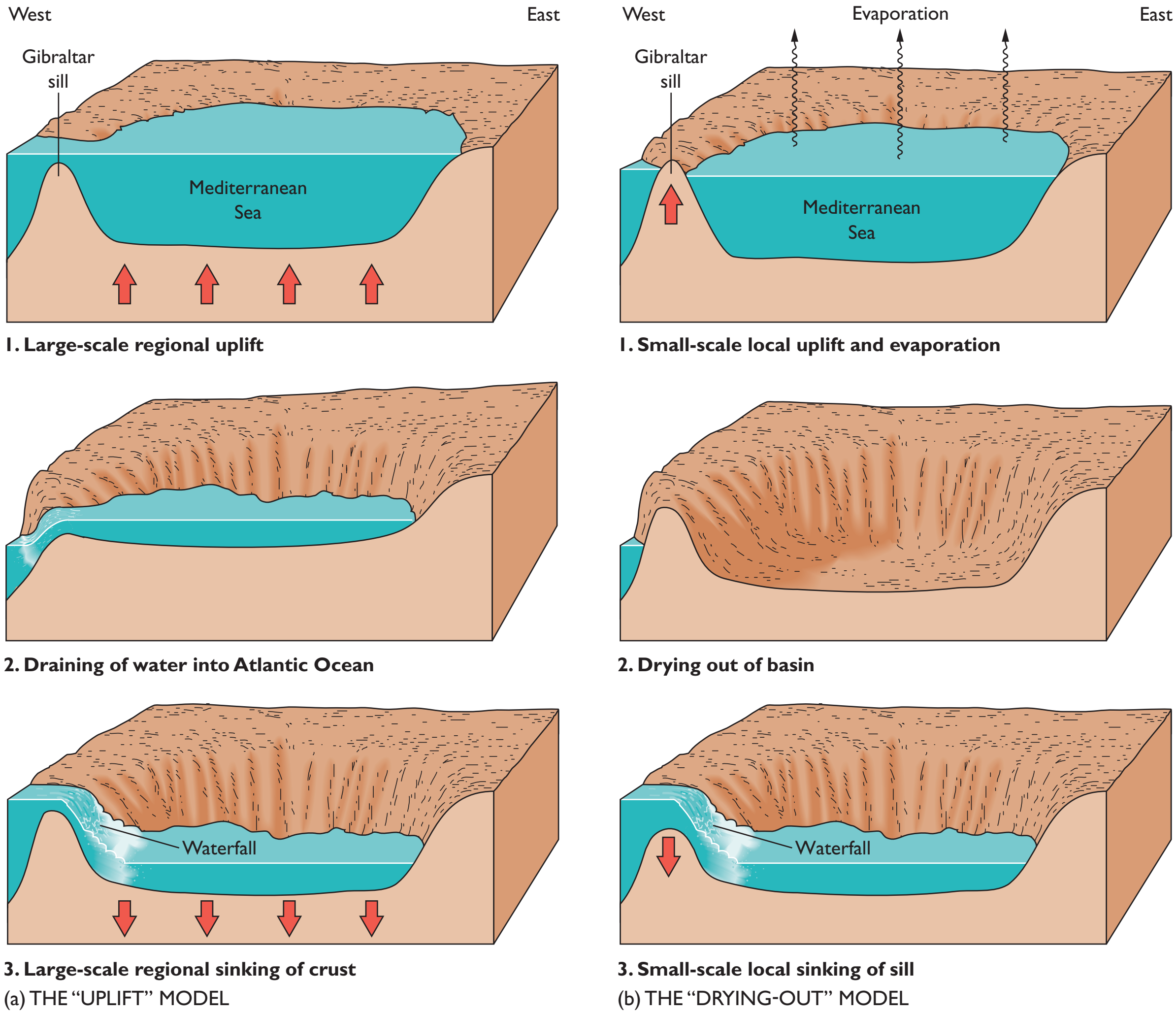

FIGURE B4–8 **Models for emptying the Mediterranean Sea.** Deep-sea drilling in the Mediterranean Sea has uncovered anhydrite and stromatolites of Miocene age. These are indicators of shallow, high-salinity conditions similar to those that exist in the present-day Arabian Gulf. These shallow-water deposits are overlain and underlain by deep-sea oozes, a peculiar stratigraphy that can be explained in two ways: (a) According to the "uplift" model, the sea bottom was thrust up thousands of meters and drained of seawater and then quickly plunged downward to abyssal depths and refilled with seawater. (b) According to the "drying-out" model, the Mediterranean Sea dried up by the evaporation of all of its water when it became separated for a while from the Atlantic Ocean and then refilled with seawater when the connection to the Atlantic was reestablished.

deeper sea floor of the eastern Mediterranean at that time was covered by a system of large lakes, similar to the Great Salt Lake in Utah, fed by spillover from the Black Sea to the east. Eventually, the Mediterranean basin refilled in a rather catastrophic fashion. Atlantic seawater cascaded down the Gibraltar Strait once its sill had dropped a bit, creating a monumental waterfall (FIGURE B4–9b). The quantity of water that poured over the sill exceeded the discharge of Niagara Falls by a factor of 1,000 (3 orders of magnitude) according to Hsü's calculations. At that rate, the Mediterranean basin would have entirely filled with Atlantic seawater in a mere 100 years!

Although this model appeals to our imaginations, it must be noted that some geologists believe that the evidence Hsü cites to support his model can be interpreted in other, less dramatic ways. The debate continues.

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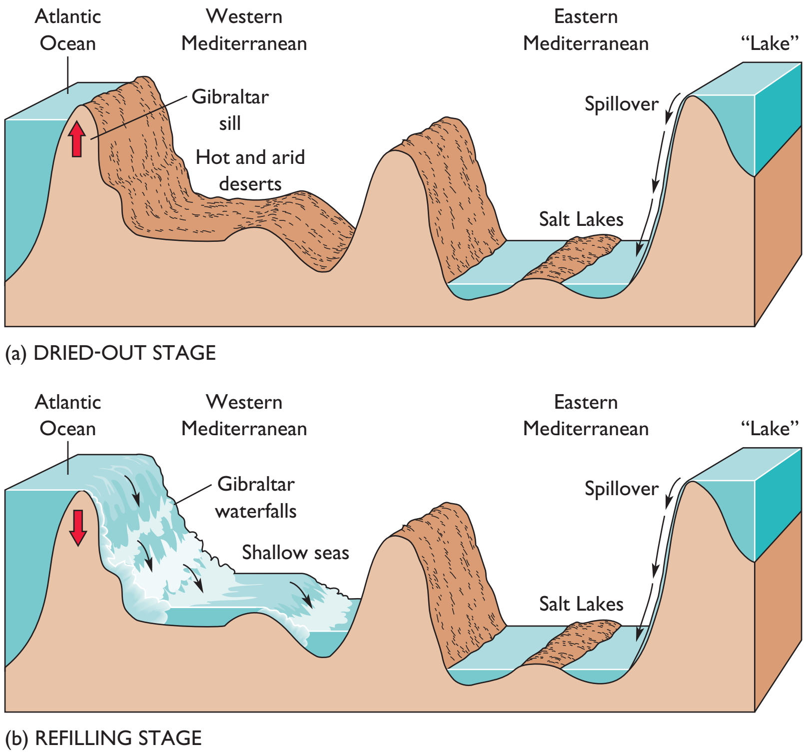

FIGURE B4–9 **The Mediterranean region during Miocene time.** (a) During the Miocene, the Mediterranean Sea apparently dried up, creating deserts and large lakes of high-salinity water. (b) Subsequently, seawater from the Atlantic Ocean cascaded down the rock face of the Gibraltar sill, refilling the entire Mediterranean basin. [Adapted from Cita, M. B., and Ryan, W. B. G. *Initial Reports of the Deep-Sea Drilling Project*, Vol. 13 (1973): 1405–1415.]

SCIENCE BY THE NUMBERS

Sedimentation Rates

Let's assume that sediment is settling on a seabed at an average rate of 5 centimeters per thousand years (5 cm/10^3 yr). Now, let's assume that this seafloor has been spreading at a rate of 4 centimeters per year. How thick will the sedimentary cover be for seafloor that is 400 kilometers down the flank of this spreading ridge? If we think a bit about this problem, we discover that to solve it all we need to know is the age of the seafloor that is located 400 kilometers from the ridge axis. The thickness of the sediment cover will be equal to the sedimentation rate times the age of the seafloor. Obviously, the older the seabed, the thicker must be its sediment cover for a given rate of deposition.

The first step is determining the age of the seafloor that is located 400 kilometers from the axis of a ridge that is spreading at a rate of 4 centimeters per year. This is similar to determining the time that has elapsed since you left home in your car, traveling at 50 mph for 100 miles. The answer is 2 hours, which you determine by dividing 100 miles (the distance traveled) by 50 miles per hour (the speed of the automobile). We do

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exactly the same calculation for the seabed problem. We divide 400 kilometers (the distance traveled) by 4 centimeters per year (the speed of seafloor spreading). In order to do that, of course, we need the same distance units. So, we must first convert kilometers into centimeters. Let's do that.

$$400 \text{ km} = 4 \times 10^2 \text{ km},$$

and

$$\left(4 \times 10^2 \cancel{\text{km}}\right)\left(10^3 \cancel{\text{m}}/\cancel{\text{km}}\right)\left(10^2 \text{ cm}/\cancel{\text{m}}\right)$$

$$= 4 \times 10^{(2+3+2)} \text{ cm} = 4 \times 10^7 \text{ cm}.$$

Now we divide the distance by the speed of seafloor spreading. (Recall the car-travel example.)

$$4 \times 10^7 \cancel{\text{cm}}/4 \cancel{\text{cm}}/\text{yr} = 10^7 \text{ yr}.$$

So, the age of the seafloor 400 kilometers away from the ridge axis is 10 million years.

The next step is determining how much sediment would have built up in 10 million years if the average sedimentation rate is 5 centimeters per 10^3 years. This is a simple matter. We multiply the sedimentation rate by the time over which it operates. Think about it this way. If you're building up a brick wall at 1 foot per hour, how high will the wall be after 3 hours? Obviously, 3 feet, which you determined by multiplying 1 foot per hour by 3 hours. We'll do the same for our sediment-thickness problem.

$$\left(5 \text{ cm}/10^3 \cancel{\text{yr}}\right)\left(10^7 \cancel{\text{yr}}\right) = 5 \times 10^{(7-3)} \text{ cm}$$

$$5 \times 10^4 \text{ cm} = 50{,}000 \text{ cm}.$$

Now we convert centimeters into kilometers as follows:

$$\left(5 \times 10^4 \cancel{\text{cm}}\right)\left(1 \cancel{\text{m}}/10^2 \cancel{\text{cm}}\right)\left(1 \text{km}/10^3 \cancel{\text{m}}\right)$$

$$= 5 \times 10^{(4-2-3)} \text{ km} = 5 \times 10^{(-1)} \text{ km}.$$

The sediment thickness 400 kilometers from this ridge is

$$5 \times 10^{(-1)} \text{ km} = 0.5 \text{ km} = 500 \text{ m}.$$

Study Guide

key concepts

1. Deposition of sediment is affected by the type, size, chemical composition, and quantity of the sediment supply; by the energy conditions at the site of deposition; and by fluctuations of sea level. Typically, coarse sediments (sand) signify high-energy conditions, fine sediments (mud) signify low-energy conditions.
2. Based on origin, ocean sediments include: *terrigenous* (eroded from land), *biogenous* (derived from organisms), *hydrogenous* (precipitated from water), *volcanogenous* (derived from volcanoes), and *cosmogenous* (fallout from outer space).
3. Shelf sediments vary in composition with latitude (Figure 4–4a). In the tropics and subtropics, they consist of *biogenous* deposits composed of the hard calcium carbonate remains of organisms. Shelf sediments of the midlatitudes consist mainly of river-supplied quartz and feldspar sands, *terrigenous sediment*, and those of the high latitudes of unsorted glacial debris.
4. The bulk of the sediment of continental shelves is *relict* (Figures 4–3b and 4–4b), having been deposited during the latest glacial event of the Pleistocene epoch when sea level was much lower than it is at present. Relict sediments are not in equilibrium with the present-day water depths and bottom-energy conditions of the shelves, but, rather, reflect an old pattern of deposition.
5. Over intervals of more than a million years, the evolution of continental shelves depends

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on the plate-tectonic setting. *Atlantic-type margins* develop along the edges of continents that are located far from the tectonic effects of plate boundaries and have a long history of passive sedimentation (Figures 4–5d and 4–6). In contrast, *Pacific-type margins* are dominated by a subduction zone and are built up of highly deformed sedimentary beds that may be interlayered with lavas and volcanogenous sediment (Figures 4–5c and 4–7).

6. During the past million years, continental shelves have been periodically uncovered and covered by the ocean as glaciers advanced and retreated on land (Figures 4–2b and 4–3a). Low stands of sea level resulted in glaciers and rivers cutting into the exposed shelf surface, and *slumping* on the continental slope (Figure 4–10a), and the flow of *turbidity currents* down submarine canyons increased the amount of sand and mud transported to the deep sea (Figures 4–10b, c, and d). Exposure caused by the drop in sea level during glacial periods caused widespread killing of coral reefs in the low latitudes.
7. Modern deposits on shelves have been influenced by waves and tidal currents. Most of the coarse sediment of continental shelves is moved during storms, when the combination of large waves and tidal currents create high-energy conditions along the bottom.
8. *Terrigenous sediments* are derived from the weathering and erosion of rocks on land. This material blankets the continental shelves, largely as relict material, and is transported to the deep sea by *ice rafting* (Figures 4–12) and by *bulk emplacement:* slumping, debris flows, and turbidity currents (Figure 4–10).
9. *Pelagic sediments* (Table 4–2) are fine-grained materials that have settled out of the water, particle by particle. *Pelagic clay* is largely derived from the weathering of rocks on land. *Biogenous sediments,* which are composed of the skeletal remains of microscopic organisms, are divided into *calcareous oozes* and *siliceous oozes* based on the chemical composition of the dominant fossils (Figure 4–14).
10. *Hydrogenous sediments* (such as *ferromanganese nodules*) are geochemical and biochemical deposits that form at the sea bottom (Figure 4–15) and are not transported to the site of deposition.
11. In the deep sea, terrigenous deposits tend to collect near the mouths of submarine canyons and are brought there by turbidity currents that flow down the canyons. Siliceous oozes form distinct bands of deposits at the equator and at the polar latitudes. Carbonate oozes accumulate in the center of ocean basins in water depths above the *carbonate compensation depth* (CCD). Pelagic clays are deposited in the most remote and deepest regions of the sea below the CCD, and away from areas having high biological productivity. Figure 4–16 shows the distribution of deep-sea deposits.
12. Because the crests of ocean-spreading ridges lie above the CCD, the first sediment to cover the basalt crust is carbonate ooze, which thickens with distance down the flank of the ridge. Once the seafloor of a spreading ridge sinks below the CCD, pelagic clays accumulate and form a second sedimentary layer that covers the basal ooze (Figure 4–17). If the spreading ocean floor crosses the equator, a third layer of ooze is produced by the tremendous fallout of biogenous debris from above. Then, beyond the zone of high surface productivity, a fourth layer composed of mud accumulates (Figure 4–18c).

key words*

accretionary prism (108)
Atlantic-type margin (106)
biogenous sediment (92)
bulk emplacement (110)
calcareous ooze (117)
carbonate compensation depth (CCD) (117)
diatoms (117)
ferromanganese nodule (118)
foraminifera (117)
glacial-marine sediment (115)
graded bedding (112)
hemipelagic sediment (115)
hydrogenous sediment (92, 110)
hydrothermal vents (118)
ice rafting (115)
limestone (110, 120)
Pacific-type margin (108)
pelagic clay (115)
pelagic sediment (113, 115)
phosphorite (120)
radiolaria (117)
relict sediment (97)
sandstone (120)
shale (120)
siliceous ooze (117)
slumps (110)
stratigraphy (122)
terrigenous sediment (92)
till (100)
turbidite (112)
turbidity current (112)
volcanogenous sediment (92)

*Numbers in parentheses refer to pages.

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Review of Basic Concepts

1. What is sediment, and what is the difference between biogenous and terrigenous sediment?
2. How do geologists determine that sea level has fluctuated widely in the geologic past?
3. What is relict sediment, and why is this the dominant type of material that blankets the continental shelves of the world?
4. What distinguishes an Atlantic-type from a Pacific-type continental margin, and how are these distinctions reflected in their respective sedimentary deposits?
5. What factors control the grain size of shelf sands and muds, tillites, turbidites, and oozes?
6. Distinguish clearly between bulk emplacement and pelagic sedimentation. What specific kinds of sediment do each supply to the deep sea?
7. What grain sizes, textures, compositions, or other properties would allow you to distinguish turbidites from glacial-marine sediment, diatom ooze from foram ooze, and diatom ooze from red clay?
8. What is the carbonate compensation depth (CCD), and how does it affect deep-sea sedimentation? Trace where the CCD touches the seafloor on the map in Figure 4–16.

Critical-Thinking Essays

1. Why is there relict sediment on the continental shelves of the world? Is there relict sediment on the bottom of the deep sea? Why or why not? (Hint: Think carefully about the definition of the term.)
2. If you were sampling biogenic ooze from water depths of about 5.5 kilometers, would it be composed of silica or calcium carbonate? Explain. Would you be most likely to sample this ooze from the deep-sea bottom of the low, middle, or high latitudes? Why?
3. Why are manganese nodules common on seabeds covered by pelagic clay but rare on seabeds covered by oozes?
4. Examine the maps in Figures 4–16a and 4–16b, and account for the distribution and accumulation rates of bottom sediments in the Indian Ocean.
5. Where in the deep sea are sediments thickest, thinnest, youngest, and oldest? Explain your reasoning.
6. How would you determine from stratigraphy whether or not a section of the deep seafloor and its sediments had drifted across the equator?
7. Examine Table 4–2. Why is the dominant pelagic sediment in the Pacific Ocean composed of pelagic clay and the dominant pelagic sediment in the Atlantic Ocean composed of foraminiferal ooze?

Discovering with Numbers

1. Consult Figure 4–1. Estimate the current velocity required to erode gravel and to erode clay. Given the great difference in their size, why are their erosion velocities similar?
2. Consult Figure 4–2b. Calculate the rate of sea level rise in meters per year and in meters per 10^3 years during the past 5,000 years, and between 5,000 and 10,000 years ago.
3. Considering your response in Question 2 above, estimate in years when a city that lies 2 meters above present-day sea level is likely to be flooded. What critical assumptions must you make in order to make the estimate?
4. Convert a sedimentation rate of 5 cm/1,000 yrs into centimeters per year (cm/yr) and inches per year (in./yr).
5. Assuming a sedimentation rate of 5 centimeters per 1,000 years and a seafloor spreading rate of 5 centimeters per year, calculate how far you would have to travel from the crest of a spreading ridge to encounter sediment that is 100 meters thick.
6. Assume that you go to the spot calculated in Question 4 above and you discover that the sediment is not 100 meters thick here, but is 185 meters thick! Assume that the spreading rate in Question 4 is correct and recalculate the sedimentation rate. Now assume that the sedimentation rate in Question 4 is correct and recalculate the spreading rate.
7. Assume that plate A is subducting plate B, i.e., plate A is overriding plate B. Plate A is spreading toward the subduction zone at 10 cm/yr, plate B at 5 cm/yr. Calculate in kilometers the amount plate B is subducted by plate A in 10^6 years and 2.5×10^8 years.

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