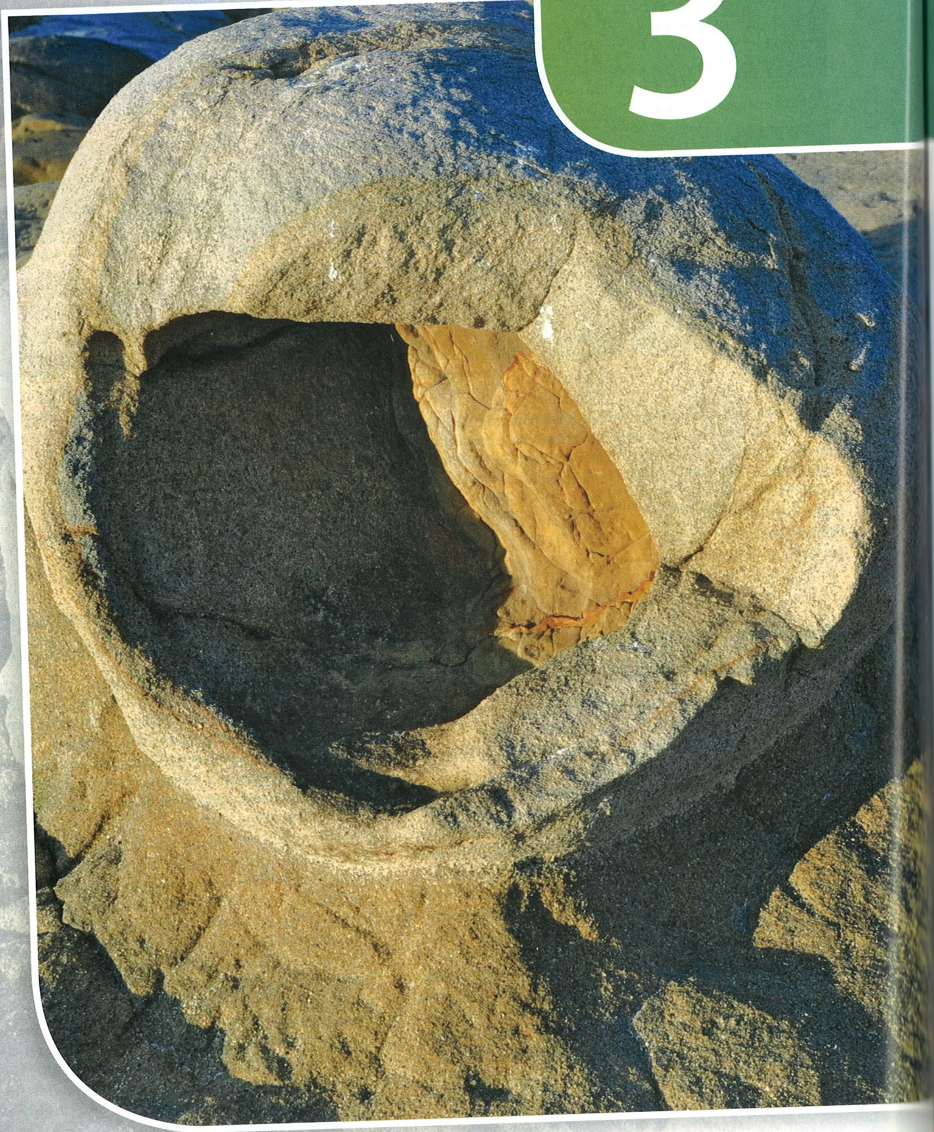
CHAPTER

3

The Sea Floor and Its Sediments



Learning Outcomes

After studying the information in this chapter students should be able to:

1. *review* the evolution of methods to measure ocean depth from the time of the ancient Greeks to the present,
2. *construct* a simple cross section of an ocean basin, including both passive and active continental margins,
3. *discuss* the formation of atolls,
4. *sketch* the location of ocean ridges and trenches,
5. *explain* three different ways to classify sediments,
6. *list* the organisms that produce the majority of calcareous and siliceous sedimentary particles,
7. *identify* where biogenous and lithogenous sediments are dominant on the sea floor,
8. *define* isotopes and describe how they can be used with marine sediments as historical records, and
9. *list* multiple seabed resources and *appraise* the extent to which they are currently being recovered.

CHAPTER OUTLINE

A large eroded sandstone boulder on the coast.

Early mariners and scholars believed that the oceans were large basins or depressions in Earth's crust, but they did not conceive that these basins held features that were as magnificent as the mountain chains, deep valleys, and great canyons of the land. As maps became more detailed and as ocean travel and commerce increased, measurement of water depths and recording of seafloor features in shallower regions became necessary to maintain safe travel and ocean commerce. The secrets of the deeper oceanic areas had to wait for hundreds of years until the technology of the late twentieth century made it relatively easy to map and sample the sea floor. It was only then that large numbers of survey vessels accumulated sufficient data to provide the details of this hidden terrain.

What we know about the sea floor and its covering of sediments comes almost entirely from observations by surface ships; more recently, submersibles, robotic devices, and satellites have added to our knowledge. Some areas of the sea floor have been measured in great detail; charts of other areas have been made from scanty data. The demand for more measurements to describe and explain the features of the sea floor continues.

In this chapter, we survey the world's ocean floors and discuss their topography and geology. We examine the sources, types, and sampling of sediments and discuss seabed mineral resources.

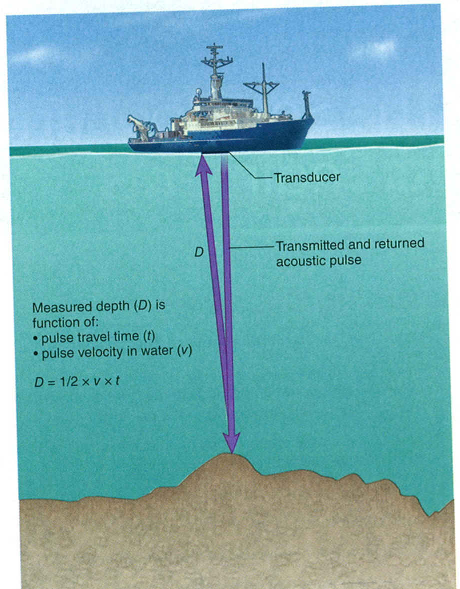


Figure 3.1 Measuring water depth (*D*) with a precision depth recorder (PDR), or echo sounder. Sound travels a total distance of twice the water depth. Depth can be calculated knowing the speed of sound in water (*v*) and the time (*t*) it takes to hear the echo, or reflection, of the sound off the bottom.

3.1 Measuring the Depths

In about 85 B.C., a Greek geographer named Posidonius set sail, curious about the depth of the ocean. He directed his crew to sail to the middle of the Mediterranean Sea, where they eased a large rock attached to a long rope over the side. They lowered it nearly 2 km (1.2 mi) before it hit bottom and answered Posidonius's question. Crude as this method was, it continued with minor modifications as the means of obtaining **soundings**, or depth measurements, for the next 2000 years.

An early modification made by nineteenth-century surveyors was the use of hemp line or rope with a greased lead weight at its end. This line was marked in equal distances (usually **fathoms**; a fathom is the length between a person's fully outstretched hands, standardized at 6 ft). The change in line tension when the weight touched bottom indicated depth, and the particles from the bottom adhering to the grease confirmed the contact and brought a bottom sample to the surface. This method was quite satisfactory in shallow water, and the experienced captain used the properties of the bottom sample to aid in navigation, particularly at night or in heavy fog.

Later, piano wire with a cannonball attached was used in deep water. These nineteenth-century surveyors used a mechanical sounding machine that allowed the rope or wire to free-fall. Using a clock, they carefully timed the rate at which premeasured lengths of the sounding line paid out over the side of the ship. This rate was a constant as long as the weight was free-falling toward the bottom but when the rate of payout decreased abruptly, they knew the weight had hit the bottom. In this way, they were able to obtain reasonably precise soundings in deep water even when the ship was moving at a slow speed. The time required (eight to ten hours) to let the weight free-fall to the bottom and then winch it back up in deep water was so great that by 1895, only about 7000 measurements had been made in water deeper than 2000 m (6600 ft) and only 550 measurements had been made of depths greater than 9000 m (29,500 ft).

It was not until the 1920s, when acoustic sounding equipment was invented, that deep-sea depth measurements became routine. The **echo sounder**, or **depth recorder**, which measures the time required for a sound pulse to leave the surface vessel, reflect off the bottom, and return, allows continuous measurements to be made easily and quickly when a ship is underway (fig. 3.1). The behavior of sound in seawater and its uses as an oceanographic tool are discussed in chapter 4. A trace from a

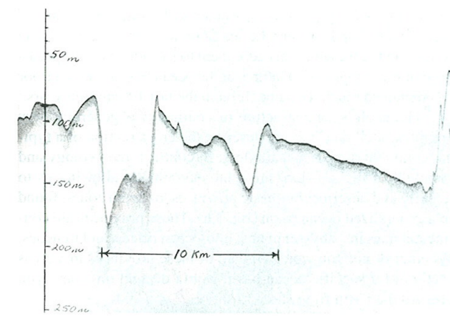


Figure 3.2 Depth recorder trace. A sound pulse reflected from the ocean floor traces a depth profile as the ship sails a steady course. The horizontal scale depends on ship speed.

depth recorder is shown in figure 3.2. (See figs. 4.18 and 4.19 for an illustration of how sound can be used to measure water depth and a picture of a precision depth recorder.)

In 1925, the German vessel *Meteor* made the first large-scale use of an echo sounder on a deep-sea oceanographic research cruise and detected the Mid-Atlantic Ridge for the first time. After this expedition, depth measurements gradually accumulated at an ever-increasing rate. As the acoustic equipment improved and was used more frequently, knowledge of the ocean floor's bathymetry expanded and improved, culminating in the 1950s with the first detailed mapping of the mid-ocean ridge and trench systems.

Today, a wide variety of methods are used to obtain even more detailed seafloor bathymetry at scales that range from centimeters (inches) to thousands of kilometers (thousands of miles) (fig. 3.3). The specific technique used depends on the amount of time that can be spent, the scale of the feature that is being examined, and the amount of detail that is required. When necessary, direct observation of small-scale structures is possible with the use of staffed submersibles and remotely operated vehicles (ROVs) carrying video cameras. These images can be transmitted to surface ships and relayed by satellite anywhere in the world in real time. Investigations by staffed submersibles or ROVs provide great detail, but they typically cover very small areas and are both time-consuming and expensive for the amount of sea floor surveyed. On large scales of tens or hundreds of square kilometers, sophisticated multibeam sonar systems can rapidly map extensive regions at relatively low cost with great accuracy.

Very-large-scale seafloor surveys use satellite measurements of changes in sea surface elevation caused by changes in Earth's gravity field due to seafloor bathymetry. These changes in sea surface elevation can be detected by radar altimeters that measure the distance between the satellite and the sea surface (fig. 3.4). The sea surface is not flat even when it is perfectly calm. Changes in gravity caused by seafloor topography create gently sloping hills and valleys in the sea surface. The excess mass of features such as seamounts and ridges creates a gravitational attraction that draws water toward them, resulting in a higher elevation of the sea surface. Conversely, the deficit of mass along deep-ocean trenches, and subsequent weaker gravitational attraction, results in a depression of the sea surface as water is drawn away toward surrounding areas with greater gravitational attraction. Sea level over large seamounts is elevated by as much as 5 m (16 ft) and over ocean ridges by about 10 m (33 ft); it is depressed over trenches by about 25–30 m (80–100 ft). These changes in elevation occur over tens to hundreds of kilometers, so the slopes are very gentle. The sea surface is always perpendicular to the local direction of gravity, so precise measurements of the slope of the surface can be used to determine the direction and magnitude of the gravitational field at any point. Because these changes in gravity are related to seafloor topography, it is possible to use them to reconstruct a bathymetry that produces the observed variations in sea surface topography (fig. 3.5). Tides, currents, and changes in atmospheric pressure can cause undulations of more than a meter (3 feet) in the ocean surface. These effects are filtered out

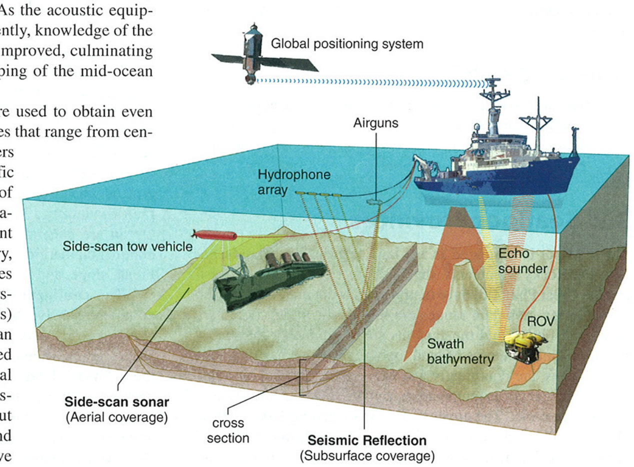


Figure 3.3 Mapping the sea floor using sound waves can be done with many different instruments depending on the level of detail required and whether the goal is to simply produce a map of the sea floor or to also map layers of sediment and rock beneath the sea floor. The various methods used differ in the intensity, frequency, and swath of the emitted sound signal, and in the method used to detect and record the echo.

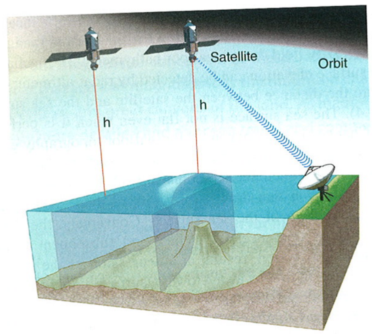


Figure 3.4 Satellite altimeters determine the elevation of the ocean surface by measuring the precise travel time of radar signals. The orbit of the satellite is known with a high degree of accuracy. The gravitational attraction of a seamount causes water to be drawn toward it, increasing the elevation of the sea surface above it.

to produce the bathymetric details. Bathymetric features with "footprints," or horizontal dimensions as small as about 10 km (6.2 mi), can be resolved with satellite altimetry data. Satellite maps are particularly valuable in the Southern Ocean, where the weather and sea conditions are frequently bad and it is difficult to conduct general bathymetric surveys to locate areas of scientific interest.

QUICK REVIEW

1. Why do oceanographers use sound rather than light to measure ocean depths?
2. Does satellite altimetry measure ocean depth directly or indirectly? Explain your answer.

3.2 Seafloor Provinces

The sea floor is as rugged as any land surface. The Grand Canyon, the Rocky Mountains, the desert mesas in the Southwest, and the Great Plains all have their undersea counterparts. In fact, the undersea mountain ranges are longer, the valley floors are wider and flatter, and the canyons are often deeper than those found on land. Features of land topography, such as mountains and canyons, are continually and aggressively eroded by wind, water, ice, changes in temperature, and the chemical alteration of minerals in rocks. The erosion of seafloor bathymetry is generally slow. Physical weathering occurs primarily by waves and currents, and chemical erosion occurs by the dissolution of minerals. More rapid erosion generally occurs closer to coastlines.

The most important agents of physical change on the deep-sea floor are the gradual burial of features by a constant rain of sediments falling from above, and volcanism associated with the mid-ocean ridge system, hotspots, island arcs, and active seamounts and abyssal hills. Movements of Earth's crust can displace features and fracture the sea floor, and the weight of some islands and seamounts can cause them to subside, but the appearance of the bathymetric features of the ocean basins and sea floor has remained much the same through the last 100 million years.

The geology and structure of continents is generally very complex, making efforts to describe the cross section of a typical continent almost meaningless. In contrast, the geology and structure of the sea floor are relatively simple, allowing us to identify and describe four basic provinces, or depth zones, found in a generalized ocean basin (fig. 3.6). These provinces are continental margins, abyssal plains, mid-ocean ridges, and trenches. Whether or not you would find all of these provinces in a cross section of a specific ocean basin would depend on where you crossed the basin from shore to shore.

Continental Margins and Submarine Canyons

The edges of the landmasses below the ocean surface and the steep slopes that descend to the sea floor are known as the **continental margin**. There are two basic types of continental margins: passive margins and active margins (fig. 3.6). **Passive margins** are found where the continent-ocean transition is not a plate boundary. The transition from continental crust to oceanic crust occurs within a single plate. Passive margins have little seismic or volcanic activity and they tend to be relatively wide. They form after continents are rifted apart, creating a new and growing ocean basin between them. The continental margins in the Atlantic Ocean are of this type. **Active margins** are found where the continent-ocean transition is a plate boundary. In moving from continental crust to oceanic crust, you move from one tectonic plate to another. Active margins are often associated with earthquakes and volcanism and are often relatively narrow. Most active margins are associated with plate convergence and subduction of oceanic lithosphere beneath a continent, creating an ocean trench (review fig. 2.30*b*). Some active margins are created when two plates slide past each other along a transform fault (review fig. 2.29). The continental margins in the Pacific Ocean are generally active margins. The general model of a continental margin consists of four parts: continental shelf, shelf break, slope, and rise (fig. 3.7).

The **continental shelf** lies at the edge of the continent; continental shelves are the nearly flat borders of varying widths that slope very gently away from the shoreline. The continental shelves are geologically part of the continental crust; they are the submerged seaward edges of the continents. Shelf widths average about 65 km (40 mi) but are typically much narrower along active margins than along passive margins. In figure 3.5 you can see the striking difference between the narrow shelf of the active continental margin along the Pacific coast of South America and the broad shelf of the passive continental margin along South America's Atlantic coast. The width of the continental shelf can be as much as 1500 km (930 mi). Water depth at the outer edge of the continental shelf varies, but on average it is about 130 m (430 ft).

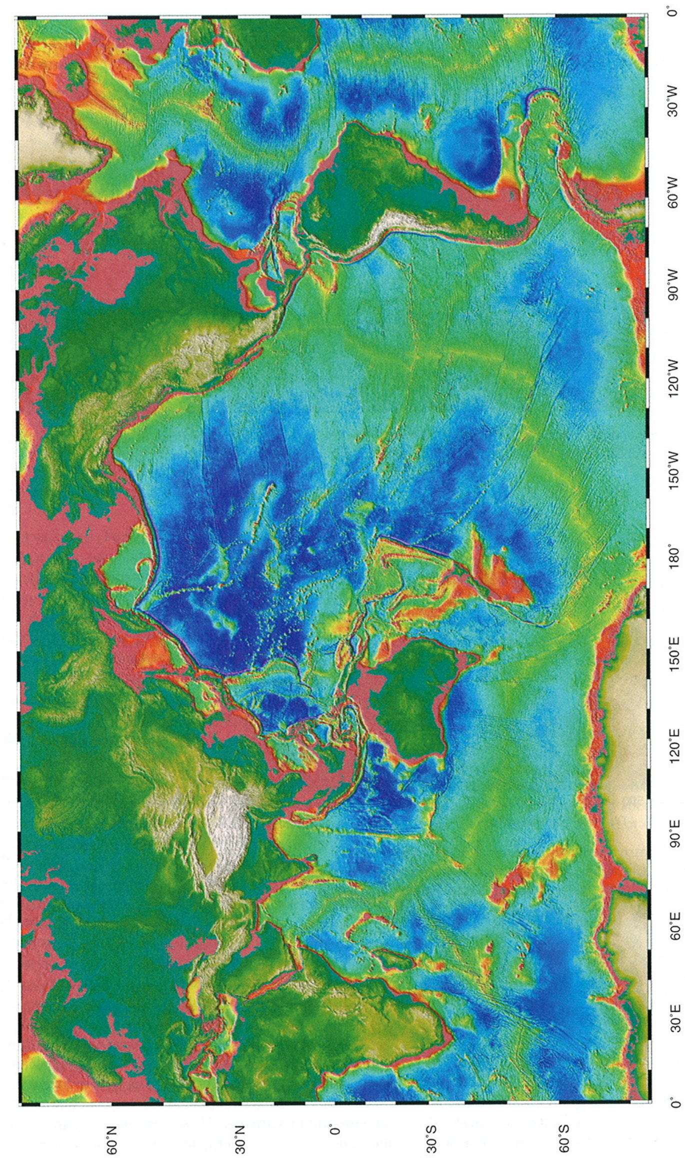


Figure 3.5 Color-shaded relief image of the bathymetry of the world's ocean basins modeled from marine gravity anomalies mapped by satellite altimetry and checked against ship depth soundings.

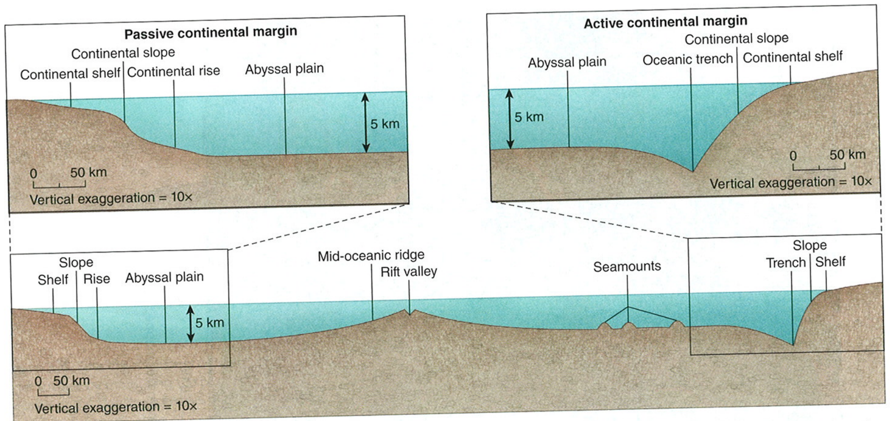


Figure 3.6 A cross section of a generalized ocean basin consists of four basic provinces: the continental margin (of some sort), the abyssal plain, a ridge, and a trench. Specific ocean basins can have different types of continental margins and may or may not have a trench.

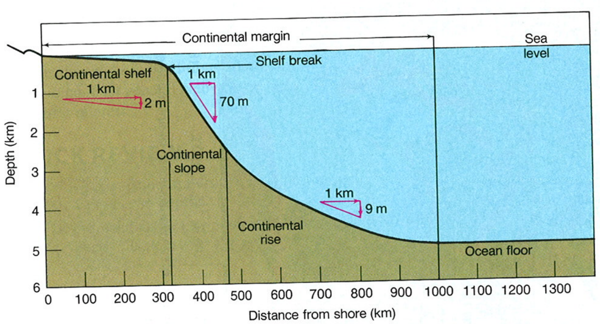


Figure 3.7 A typical profile of a passive continental margin. Notice both the vertical and horizontal extent of each subdivision. The average slope is indicated for the continental shelf, slope, and rise. The vertical scale is 100 times greater than the horizontal scale (V.E. = 100 times).

During past ages, the shelves have been covered and uncovered by fluctuations in sea level. During the glacial ages of the Pleistocene epoch, a number of short-term changes occurred in sea level, some of which were greater than 120 m (400 ft). When sea level was low, erosion deepened valleys, waves eroded previously submerged land, and rivers left sediments far out on the shelf. When the glacial ice melted, these areas were flooded, and sediments built up in areas closer to the new shore. At present, although submerged, these areas still show the scars of old riverbeds and glaciers acquired when the land was above water. Today, some continental shelves are covered with thick deposits of silt, sand, and mud sediments derived from the land: examples are offshore from the mouths of the Mississippi and Amazon Rivers, where large amounts of such sediments are deposited annually. Other shelves are bare of sediments, such as where the fast-moving Florida Current sweeps the tip of Florida, carrying the sediments northward to the deeper water of the Atlantic Ocean.

The **continental shelf break** is an abrupt change in the slope of the sea floor that occurs at the outer edge of the continental shelf. This marks the point at which there is a rapid increase in depth with distance from the coast.

The **continental slope** dips relatively steeply down to the ocean basin floor. The angle and extent of the slope vary from location to location. The slope can be short and steep, dropping to depths of around 3000 m (10,000 ft) along passive margins (as in fig. 3.7), or it might drop as far as 8000 m (26,000 ft) into a deep-ocean trench along an active margin (for example, off the western coast of South America, where the narrow continental shelf is bordered by the Peru-Chile Trench). The continental slope may show rocky outcroppings and be relatively bare of sediments because of its steepness, tectonic activity, or a low supply of sediment from land.

The most outstanding features of the continental slopes are **submarine canyons**. These canyons sometimes extend up, into, and across the continental shelf. A submarine canyon is

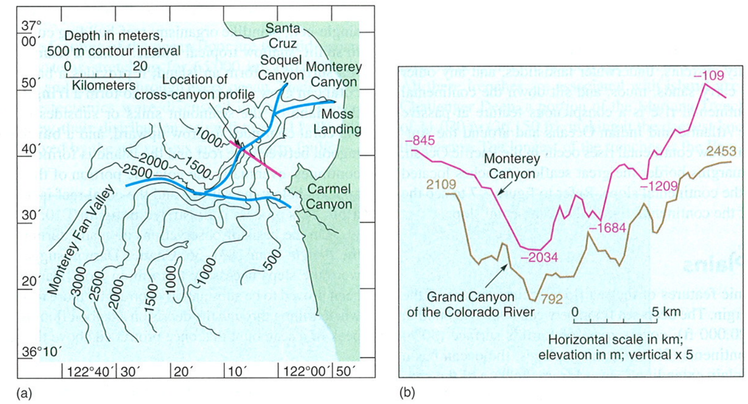


Figure 3.8 (a) Depth contours depict three submarine canyons off the California coast as they cut across the continental slope and continental shelf. The axes of the canyons, which merge seaward, are indicated by the *blue line*. (b) Cross-canyon profile, along the *red line* in (a), of the Monterey Canyon. Compare this profile to that of the Grand Canyon drawn to the same scale.

steep-sided and has a V-shaped cross section, with tributaries similar to those of river-cut canyons on land. Figure 3.8*a* shows the Monterey and Carmel Canyons off the coast of California. Figure 3.8*a* is a bathymetric chart; figure 3.8*b* compares the profile of the Monterey Canyon with the profile of the Grand Canyon of the Colorado River.

Many of these submarine canyons are associated with existing river systems on land and were apparently cut into the shelf during periods of low sea level, when the glaciers advanced and the rivers flowed across the continental shelves. Ripple marks on the floor of the submerged canyons and sediments fanning out at the ends of the canyons suggest that they were formed by moving flows of sediment and water called **turbidity currents**. Caused by earthquakes or the overloading of sediments on steep slopes, turbidity currents are fast-moving avalanches of mud, sand, and water that flow down the slope, eroding and picking up sediment as they gain speed. In this way, the currents erode the slope and excavate the submarine canyon. As the flow reaches the bottom, it slows and spreads, and the sediments settle. Because of their speed and turbulence, such currents can transport large quantities of materials of mixed sizes. The settling process produces graded beds of coarse material overlain (upward) by smaller particles. These graded deposits are called **turbidites**. Figure 3.9 shows a turbidite preserved in compacted seafloor sediments that have been uplifted and exposed by wave erosion. These large and occasional currents have never been directly observed, although similar but smaller and more continuous flows, such as sand falls, have been observed and photographed.

Research on turbidity currents began with laboratory experiments in the 1930s. Later analysis of a 1929 earthquake that broke transatlantic telephone and telegraph cables on the continental slope and rise off the Grand Banks of Newfoundland showed a pattern of rapid and successive cable breaks high on the continental slope, followed by a sequence of downslope breaks. These breaks were calculated to have been caused by a turbidity current that ran for 800 km (500 mi) at speeds of 40–55 km (25–35 mi) per hour. Later samples taken from the area showed a series of graded sediments at the end of the current's path. Searches of cable company records showed similar patterns of cable breaks in other parts of the world.

Figure 3.9 This beach cliff at Point Lobos, California shows ancient turbidite deposits that have been uplifted and then exposed by wave erosion. Turbidites are graded deposits, with the largest particles in the deposit at the bottom of the turbidite and the smallest at the top.

At the base of the steep continental slope may be a gentle lope formed by the accumulation of sediment. This portion of ne sea floor is the **continental rise**, made up of sediment deposed by turbidity currents, underwater landslides, and any other rocesses that carry sands, muds, and silt down the continental lope. The continental rise is a conspicuous feature at passive nargins in the Atlantic and Indian Oceans and around the Antrctic continent. Few continental rises occur in the Pacific Ocean, where active margins border the great seafloor trenches located t the base of the continental slope. Refer to figure 3.7 to see the elationship of the continental rise to the continental slope.

Abyssal Plains

The true oceanic features of the sea floor occur seaward of the continental margin. The deep-sea floor, between 4000 and 6000 m 13,000 and 20,000 ft), covers more of Earth's surface (30%) han do the continents (29%). In many places, the ocean basin floor is a vast plain extending seaward from the base of the continental slope. It is flatter than any plain on land and is known as the **abyssal plain.** The abyssal plain is formed by sediments that fall from the surface or are deposited by turbidity currents to cover the irregular topography of the oceanic crust. An area of the abyssal plain that is isolated from other areas by continental margins, ridges, and rises is known as a basin, and some basins may be divided into subbasins by ridge and rise subsections. The distribution of these basins and subbasins is shown in figure 3.5. Low ridges allow some exchange of deeper water between adjacent basins, but if the ridge is high, both the deep water and the deep-dwelling marine organisms within the basin are effectively cut off from other basins.

Abyssal hills and **seamounts** are scattered across the sea floor in all the oceans. Abyssal hills are less than 1000 m (3300 ft) high, and seamounts are steep-sided volcanoes rising abruptly and sometimes piercing the surface to become islands. Abyssal hills are probably Earth's most common topographic feature. They are found over 50% of the Atlantic sea floor and about 80% of the Pacific floor; they are also abundant in the Indian Ocean. Most abyssal hills are probably volcanic, but some may have been formed by other movements of the sea floor. Submerged, flat-topped seamounts, known as **guyots,** are found most often in the Pacific Ocean. The tops of Pacific guyots are 1000–1700 m (3300–5600 ft) below the surface; many are at the 1300 m (4300 ft) depth. Many guyots show the remains of shallow-water coral reefs and evidence of wave erosion at their summits. These features indicate that at one time they were warm-water surface features and that their flat tops are the result of wave erosion. They have since subsided owing to their weight, the accumulated rock load bearing down on the oceanic crust, and the natural subsidence of the sea floor with increasing age as the crust cools and grows in density as it moves farther away from the ridge where it formed. They have also been submerged by rising sea level during periods when glacial ice melted on land.

In the warm waters of the Atlantic, Pacific, and Indian Oceans, coral reefs and coral islands are formed in association with seamounts. Reef-building corals are warm-water animals that require a place of attachment and grow in intimate association with a single-cell, plantlike organism; reef-building corals are confined to sunlit, shallow tropical waters. When a seamount pierces the sea surface to form an island, it provides a base on which the coral can grow. The coral grows to form a **fringing reef** around the island. If the seamount sinks or subsides slowly enough, the coral continues to grow upward, and a **barrier reef** with a lagoon between the reef and the island is formed. If the process continues, eventually the volcanic portion of the seamount disappears below the surface and the coral reef is left as a ring, or **atoll.** This process is illustrated in figure 3.10.

On the basis of observations he made during the voyage of the *Beagle* from 1831–36, Charles Darwin suggested that these were the steps necessary to form an atoll. Darwin's ideas have been proved to be substantially correct by more recent expeditions when drilling through the debris on a lagoon floor found the basalt peak of a seamount that once protruded above the sea surface.

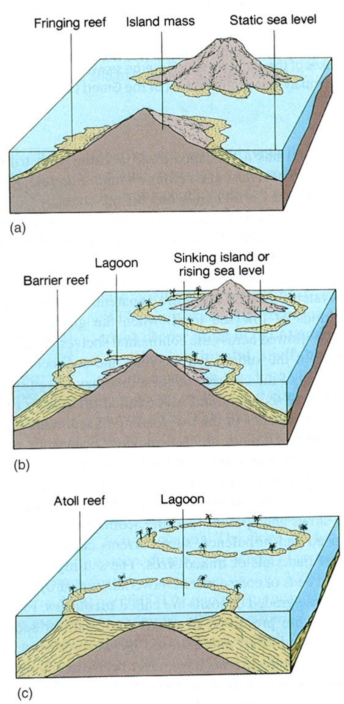


Figure 3.10 Types of coral reefs and the steps in the formation of a coral atoll shown in profile. (a) Fringing reef. (b) Barrier reef. (c) Atoll reef.

Ridges, Rises, and Trenches

The most notable features of the ocean floor are the mid-ocean ridge and rise systems stretching for 65,000 km (40,000 mi) around the world and running through every ocean. Their origin and role in plate tectonics were discussed in chapter 2 (see fig. 2.10). Review their distribution using figure 3.11. Recall also the roles played by the rift valleys and transform faults.

The relationship of the deep-sea trenches to plate tectonics was also discussed in chapter 2 (see fig. 2.10). Use figure 3.12 to trace the Japan-Kuril Trench, the Aleutian Trench, the Philippine Trench, and the deepest ocean trench, the Mariana Trench. All these trenches are associated with **island arc systems.** The Challenger Deep, a portion of the Mariana Trench, has a depth of 11,020 m (36,150 ft), making it the deepest known spot in all the oceans. The longest of the trenches is the Peru-Chile Trench,

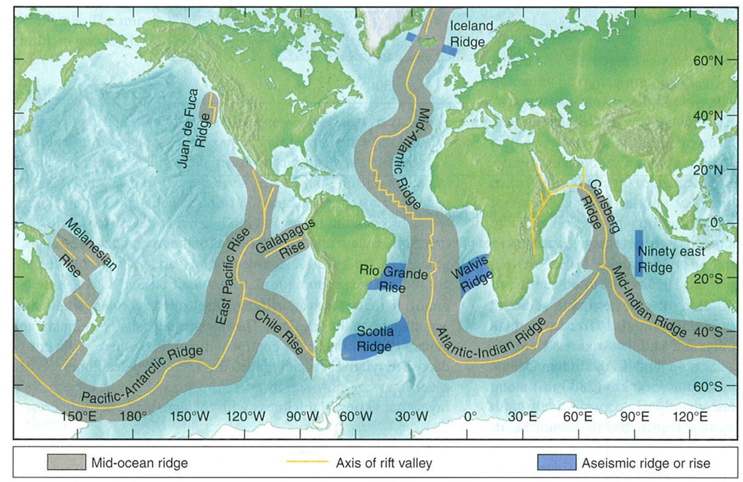


Figure 3.11 The mid-ocean ridge and rise system of divergent plate boundaries. Locations of major aseismic (no earthquakes) ridges and rises are added. Aseismic ridges and rises are elevated linear features thought to be created by hot-spot activity.

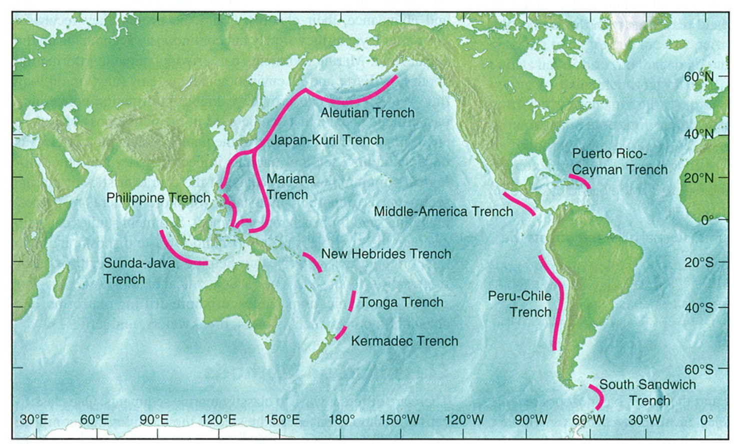


Figure 3.12 Major ocean trenches of the world. The deepest ocean depth is 11,020 m (36,150 ft), east of the Philippines in the Mariana Trench. It is known as the Challenger Deep.

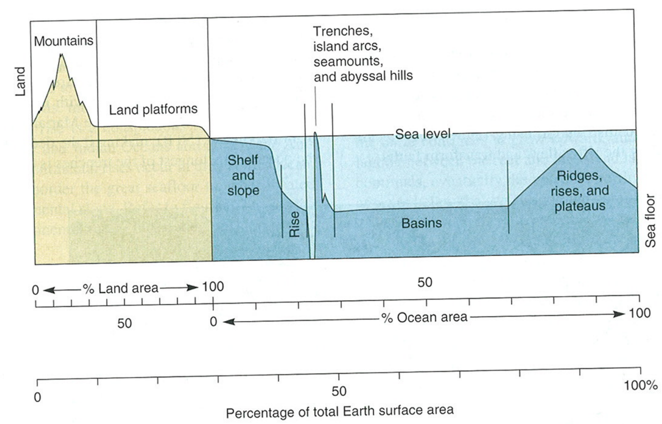


Figure 3.13 Earth's main topographic features shown as percentages of Earth's total surface and as percentages of the land and of the oceans.

stretching 5900 km (3700 mi) along the western side of South America. To the north, the Middle-America Trench borders Central America. The Peru-Chile and Middle-America Trenches are associated with volcanic chains on land. In the Indian Ocean, the great Sunda-Java Trench runs for 4500 km (2800 mi) along Indonesia. In the Atlantic, there are only two comparatively short trenches: the Puerto Rico-Cayman Trench and the South Sandwich Trench, both associated with chains of volcanic islands.

In figure 3.13, the topography of the land and the bathymetry of the sea floor are summarized as percentages of Earth's area. Compare the tectonically active areas of trenches and ridges, as well as the area of low-lying land platforms with the area of the ocean basins.

QUICK REVIEW

1. Identify the different parts of the continental margin.
2. Describe the difference between passive and active continental margins.
3. List the main provinces and features of the ocean basin floor.
4. Explain the formation of an atoll.
5. How are islands, seamounts, abyssal hills, and guyots the same? How are they different?
6. Distinguish between a ridge and a rise.
7. Where are the major deep-sea trenches?

3.3 Sediments

The margins of the continents and the ocean basin floors receive a continuous supply of particles from many sources. Whether these particles have as their origin living organisms, the land, the atmosphere, or the sea itself, they are called sediment when they accumulate on the sea floor. The thickest deposits of sediment are generally found near the continental margins, where sediment is deposited relatively rapidly; in contrast, the deep-sea floor receives a constant but slow supply of sediment that produces a thinner layer that varies in thickness with the age of the oceanic crust.

Why Study Sediments?

Oceanographers study the rate at which sediments accumulate, the distribution of sediments over the sea bottom, their sources and abundance, their chemistry, and the history they record in layer after layer as they slowly but continuously accumulate on the ocean floor. Marine sediments can provide valuable information about how Earth and its environmental systems function on long time scales. They can provide valuable information concerning past climate change that can be used in predicting possible future environmental change. More immediately, sediments can help us understand how seafloor habitats impact fisheries and other biological communities. Knowledge of the characteristics of marine sediments is important in locating offshore mineral resources, including sand for beach replenishment. Sediment studies can help evaluate the possible impact of offshore waste disposal and map offshore pollution patterns, thus helping us sustain healthy coasts. Information about sediments is also critical in identifying sites for seabed communications cables, offshore drilling platforms, and coastal structures such as piers, breakwaters, and jetties.

Classification Methods

There are different ways to classify marine sediments. One classification method is based on the size of the particles that comprise the sediment. Sediments can be described both by the range of particle sizes found in a sample and by the dominant particle size, if there is one. Another way of classifying sediments is by their geographical location—where they are found in relation to distance from the coast, for instance. The rate of deposition of sediment often varies significantly for different locations, particularly with distance from the coast. Perhaps the most common classification method for marine sediments is one that is based on the origin of the particles and their chemical composition: Where did the individual particles come from and what are they made of? Each of these methods provides specific information about sediments. The importance of that information will depend on the problem you are trying to solve by studying the sediment.

Particle Size

Sediment particles are classified by size, as indicated in table 3.1. Familiar terms such as *gravel, sand,* and *mud* are used to identify broad size ranges of large, intermediate, and small particles, respectively. Within each of these ranges, particles are further ranked to produce a more detailed scale from boulders to the very smallest clay-sized particles, which can only be seen with a microscope.

When a sediment sample is collected, it can be dried and shaken through a series of woven-mesh sieves of decreasing opening size. Material that passes through one sieve but not the next is classified by one of the sizes listed in table 3.1.

Table 3.1 Sediment Size Classifications

Descriptive Name		Diameter (mm)
Gravel	Boulder	> 256
	Cobble	64–256
	Pebble	4–64
	Granule	2–4
Sand	Very coarse	1–2
	Coarse	0.5–1
	Medium	0.25–0.5
	Fine	0.125–0.25
	Very fine	0.0625–0.125
Mud	Silt	0.0039–0.0625
	Clay	< 0.0039

A sample is said to be "well sorted" if it is nearly uniform in particle size and "poorly sorted" if it is made up of many different particle sizes. Size influences the horizontal distance a particle is transported before settling out of the water and the rate at which it sinks. In general, it takes more energy to transport large particles than it does small particles. In the coastal environment, when poorly sorted sediment is transported by wave or current action, the larger particles will settle out and be deposited first, while the smaller particles may be carried farther away from the coast and deposited elsewhere. In the open ocean, the variation in sinking rate between large and small particles has a tremendous influence on how long it takes for a particle to sink to the deep-sea floor and, hence, how far the settling particle may be transported by deep horizontal currents (table 3.2). A very fine sand-sized particle may settle to the deep-sea floor in a matter of days, where it could come to rest a short horizontal distance away from the point at the surface where it began its journey. In contrast, it may take clay-sized particles over 125 years (nearly 50,000 days) to make the same journey (see Stokes Law for small-particle settling velocity in appendix C). The speed of deep horizontal currents in the oceans is generally quite slow, but even at a speed of 5 cm (2 in) per second, a clay-sized particle could theoretically be transported around the world five times before it reached the deep-sea floor. Smaller soluble particles also have time to dissolve as they slowly sink in the deep ocean. Settling rate is also influenced by particle shape. Stokes Law assumes that the sediment particles are spherical, and the resulting calculated settling rates tend to be maximum rates. Angular grains generate small turbulent eddies that slow their rate of descent. Relatively flat particles such as clays also settle more slowly than spheres of the same density.

Scientists have puzzled over the close correlation between the particle types found in surface waters and those found almost directly below on the sea floor. This observation seems to contradict the large horizontal displacement of very slowly sinking particles due to currents in the water. Some mechanisms must be working to aggregate the tiny particles into larger particles. Scientists have observed that small particles often attract each other owing to their electrical charges. This attraction forms larger particles, which sink more rapidly. The remains of microscopic organisms known as **phytoplankton** and **zooplankton** can also be found on the sea floor. When zooplankton or phytoplankton are consumed by larger organisms, their inorganic cell walls or **tests** are expelled by these organisms in large fecal pellets that

Table 3.2 Sediment Sinking Rate and Distance Traveled

Sediment Size	Approximate Sinking Rate (m/s)	Time for a Vertical Fall of 4 km (days)	Horizontal Distance Traveled in a 5 cm/s Current (km)
Very fine sand	9.8×10^{-3}	4.7	20.4
Silt	9.8×10^{-5}	470	2040
Clay	9.8×10^{-7}	47,000	204,000

Note: The sinking rate of a particle depends on its density, shape, and diameter. These rates are based on the assumption that the particles are spherical and have a density similar to that of quartz. Estimates of the speed of deep currents vary. A conservative estimate of 5 cm/s is chosen for purposes of illustration. See appendix C for the formula for small-particle settling velocity.

sink rapidly. It is estimated that as many as 100,000 tests can be packaged into a single, large fecal pellet. This packaging of small particles into larger particles decreases the time for the remains of plankton to sink to the sea floor from years to just ten to fifteen days, minimizing their horizontal displacement by water movements. Once the fecal pellets are deposited on the bottom, breakdown of the remaining organic portion of the pellets liberates the small inorganic particles.

Location and Rates of Deposition

Marine sediments are classified as either neritic (*neritos* = of the coast) or pelagic (*pelagios* = of the sea) based on where they are found (fig. 3.14).

Neritic sediments are found near continental margins and islands and have a wide range of particle sizes. Most neritic sediments are eroded from rocks on land and transported to the coast by rivers. Once they enter the ocean, they are spread across the continental shelf and down the slope by waves, currents, and turbidity currents. The largest particles are left near coastal beaches, whereas smaller particles are transported farther from shore. Accumulation rates of neritic sediments are highly variable. In river estuaries, the rate may be more than 800,000 cm (315,000 in) per 1000 years, or 8 m (over 26 ft) per year. Each year the rivers of Asia, such as the Ganges, the Yangtse, the Yellow, and the Brahmaputra, contribute more than one-quarter of the world's land-derived marine sediments. In quiet bays, the rate may be 500 cm (about 200 in) per 1000 years, and on the continental shelves and slopes, values of 10–40 cm (about 4–16 in) per 1000 years are typical, with the flat continental shelves receiving the larger amounts. Many sediments covering the continental shelves away from river mouths were deposited thousands of years ago when sea level was lower and the shoreline was located on the shelf. Such sediments are called **relict sediments**.

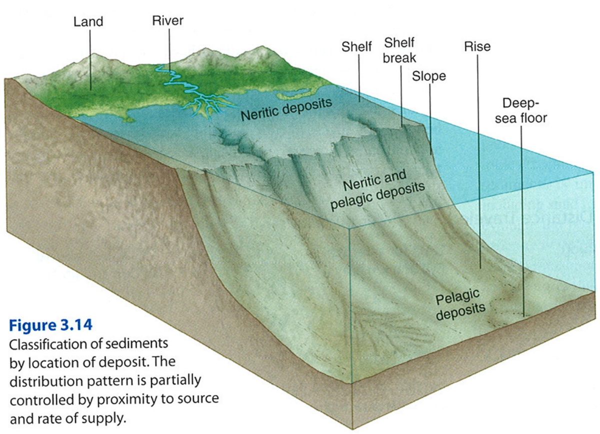


Figure 3.14 Classification of sediments by location of deposit. The distribution pattern is partially controlled by proximity to source and rate of supply.

Pelagic sediments are fine-grained and collect slowly on the deep-sea floor. The thickness of pelagic sediments is related to the length of time they have been accumulating or the age of the sea floor they cover. Consequently, their thickness tends to increase with increasing distance from mid-ocean ridges (see fig. 2.18). Accumulation rates for pelagic sediments are much slower than those of typical neritic sediments. An average accumulation rate for deep-ocean pelagic sediment is 0.5–1.0 cm (0.2–0.4 in) per 1000 years. Although deep-sea sedimentation rates are extremely slow, there has been plenty of time during Earth history for them to accumulate. The average thickness of pelagic sediments on the older areas of sea floor and the continental rises is about 500–600 m (1600–2000 ft). At a rate of 0.5 cm (0.2 in) per 1000 years, it takes only 100 million years to accumulate 500 m (1600 ft) of sediment, and the oldest sea floor is known to be roughly 200 million years old.

The actual thickness of marine sediments is controlled by a variety of factors including the age of the sea floor and its tectonic history, the nature and location of the sediment sources, and the nature of the processes that delivered the sediment to a particular location. Generally speaking, sediment thickness increases away from mid-ocean ridges as the age of the sea floor increases and maximum sediment thickness is found along continental margins where major rivers empty into the oceans (fig. 3.15).

Source and Chemistry

Marine sediments are also classified by the source of the particles that make up the sediment and may be further subdivided by their chemistry. Sedimentary particles may come from one of four different sources: preexisting rocks, marine organisms, seawater, or space.

Sediments derived from preexisting rocks are classified as **lithogenous** (*lithos* = stone, *generare* = to produce) **sediments.** These are also commonly called **terrigenous** (*terri* = land, *generare* = to produce) **sediments.** While terrigenous sediment technically includes any type of material coming off the land, such as rock fragments, wood chips, and sewage sludge, the majority of terrigenous material consists of lithogenous particles. Active volcanic islands in the ocean basins are also an important source of lithogenous sediment. Rocks on land are weathered and broken down into smaller particles by wind, water, and seasonal changes in temperature that result in freezing and thawing. The resulting particles are transported to the oceans by water, wind, ice, and gravity. Windblown dust from the continents, ash from active volcanoes, and rocks picked up by glaciers and embedded in icebergs are additional sources of lithogenous materials.

Lithogenous material can be found everywhere in the oceans. It is the dominant neritic sediment because the supply

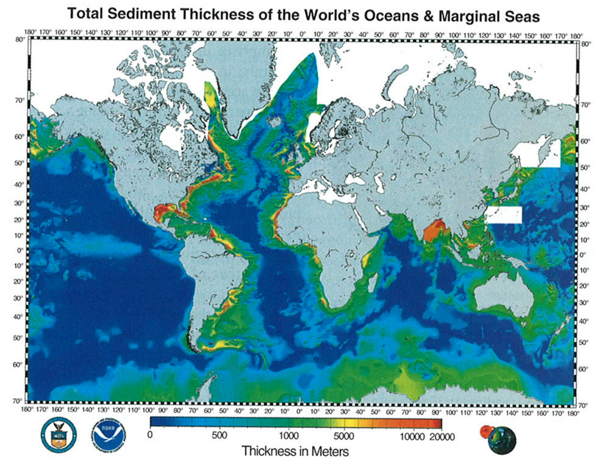


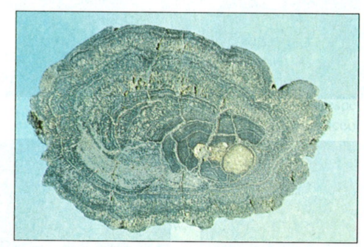
Figure 3.15 Estimated thickness of marine sediment in meters. The average thickness of deep-ocean sediments is approximately 500 m. Thinner deposits of sediments are found along the mid-ocean ridge system. Sediment thickness generally increases closer to continental margins. The thickest deposits of sediment are found near the mouths of major rivers. Source of Data: National Oceanic and Atmospheric Administration (NOAA).

of lithogenous particles from land simply overwhelms all other types of material. Pelagic lithogenous sediments on the deep-sea floor, called **abyssal clay,** are composed of at least 70% by weight clay-sized particles. Abyssal clay accumulates very slowly at rates that are generally less than 0.1 cm (0.04 in) per 1000 years. Because the accumulation rate is so slow, even a thin deposit represents a very long period of time. It is important to understand that where abyssal clay is the dominant pelagic sediment, it is only because of the lack of other types of material, not because of an increase in the supply of clay-size particles. This is generally the case in regions where there is little marine life in the surface waters above. This fine rock powder, blown out to sea by wind and swept out of the atmosphere by rain, may remain suspended in the water for many years. These clays are often rich in iron, which oxidizes in the water and turns a reddish brown color; hence, they are frequently called **red clay** (fig. 3.16*a*). The distribution of red clay is illustrated in figure 3.17.

The composition of lithogenous sediments, generally various clays and quartz, is controlled by the chemistry of the rocks they came from and their response to chemical and mechanical weathering. Most lithogenous sediments have quartz because it is one of the most abundant and stable minerals in continental rocks. Quartz is very resistant to both chemical and mechanical weathering, so it can easily be transported long distances from its source. The distribution pattern of quartz grains in the sediment can provide important information concerning changes in wind patterns and intensity through time.

Clays are abundant because they are produced by chemical weathering. Four clay minerals make up the deep-sea clays: chlorite, illite, kaolinite, and montmorillonite. The distribution of these four clays reflects different climatic and geologic conditions in the areas where and when they originated as well as along the paths they traveled before settling on the sea floor. These conditions often have a strong dependence on latitude. The warm, moist climate of low latitudes supports strong chemical weathering on land. Mechanical weathering tends to be dominant in the cold, dry climate typical of high latitudes. Chlorite is highly susceptible to chemical weathering and can be altered to form kaolinite. Consequently, chlorite is abundant in deep-sea clays at high latitudes, where chemical weathering is less effective. Kaolinite is produced in the strong chemical

(a)



(b)

Figure 3.16 (a) Manganese nodules resting on red clay photographed on deck in natural light. Nodules are 1–10 cm in diameter. (b) A cross section of a manganese nodule showing concentric layers of formation.

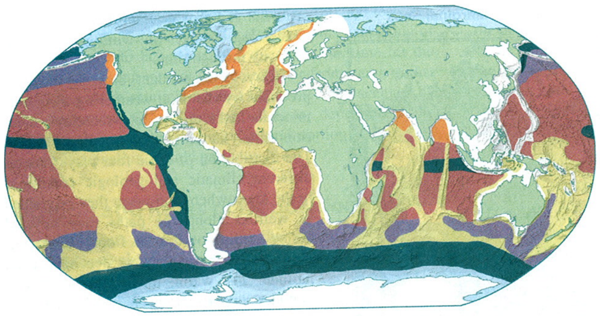


Figure 3.17 Global distribution of surficial sediments in the world ocean. Equatorial upwelling in the Pacific and ice-edge processes in the Antarctic contribute to the productivity of diatoms and, hence, the accumulation of siliceous sediments. Carbonate sediments are generally confined to shallower regions of the world ocean. Terrigenous sediments dominate near the mouths of major rivers.

weathering of minerals to form soil. It is ten times as abundant in the tropics as in polar regions, where soil-forming processes are very slow. Illite is the most widespread clay mineral. It has a clear hemispheric rather than climatic distribution. In the Southern Hemisphere, it comprises up to 20%–50% of the clay minerals; in the Northern Hemisphere, it usually accounts for more than 50% of the clay minerals. Illite forms under a variety of conditions that are not dependent on latitude, so its abundance in marine sediment depends on the degree of dilution by other clay minerals. Montmorillonite is produced by the weathering of volcanic material on land and on the sea floor. It is common in regions of low sedimentation near sources of volcanic ash. It is more abundant in the Pacific and Indian Oceans than in the Atlantic Ocean, where there is little volcanic activity along the surrounding coastlines.

Sediments derived from organisms are classified as **biogenous** (*bio* = life, *generare* = to produce) **sediments.** These may include shell and coral fragments as well as the hard skeletal parts of single-celled phytoplankton and zooplankton that live in the surface waters. Pelagic biogenous sediments are composed almost entirely of the shells, or tests, of plankton (fig. 3.18). The chemical composition of these tests is either calcareous (calcium carbonate: $CaCO_3$, as in most seashell material) or siliceous (silicon dioxide: SiO_2, clear and hard). If pelagic sediments are more than 30% biogenous material by weight, the sediment is called an **ooze;** specifically, either a **calcareous ooze** or **siliceous ooze,** depending on the chemical composition of the majority of the tests. The distribution of calcareous and siliceous oozes on the sea floor is related to the supply of organisms in the overlying water, the rate at which the tests dissolve as they descend, the depth at which they are deposited, and dilution with other sediment types (see fig. 3.17).

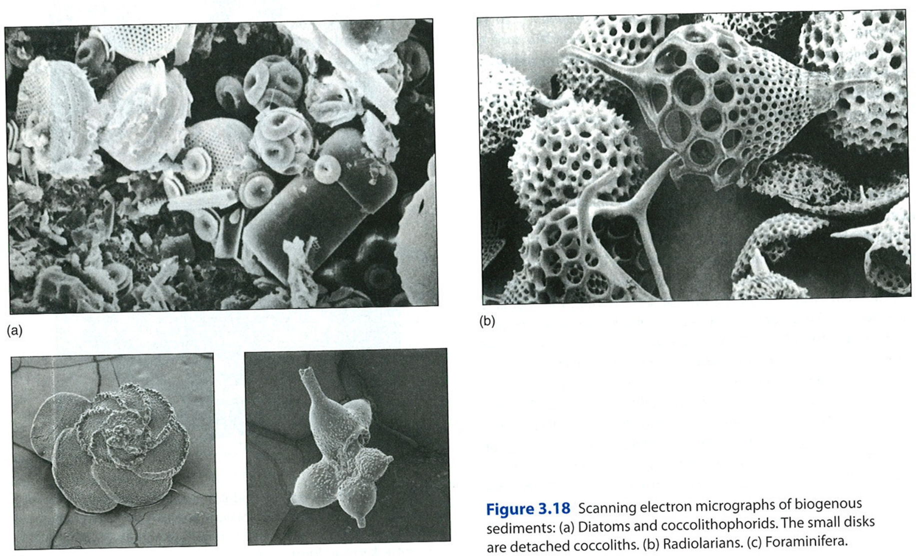


Figure 3.18 Scanning electron micrographs of biogenous sediments: (a) Diatoms and coccolithophorids. The small disks are detached coccoliths. (b) Radiolarians. (c) Foraminifera.

Calcareous tests are created by a group of phytoplankton called **coccolithophorids** (covered with calcareous plates called **coccoliths**), snails called **pteropods,** and amoeba-like animals called **foraminifera** (fig. 3.18*a* and *c*). Most coccoliths are smaller than 20 μm. Pteropod tests range from a few millimeters to 1 cm in size, while foraminifera tests range from about 30 μm to 1 mm. These deposits are often named for their principal constituent: coccolithophorid ooze, pteropod ooze, or foraminiferan ooze. Calcareous oozes are the dominant pelagic sediments (see fig. 3.17). The dissolution, or destruction, rate of calcium carbonate varies with depth and temperature and is different in different ocean basins. Calcium carbonate generally dissolves more rapidly in cold, deep water, which characteristically has a higher concentration of CO_2 and is slightly more acidic (this is discussed in detail in the sections on the pH of seawater and dissolved gas in chapter 5). The depth at which calcareous skeletal material first begins to dissolve is called the **lysocline.** Beneath the lysocline, seawater becomes undersaturated in dissolved calcium carbonate and there is a progressive decrease in the amount of calcareous material preserved in the sediment (fig. 3.19). The depth at which the amount of calcareous material preserved falls below 20% of the total sediment is called the **carbonate compensation depth (CCD).** The CCD is also commonly defined as the depth at which the rate of accumulation of calcium carbonate is equal to the rate at which it is dissolved. The lysocline and the CCD will differ in depth depending on the rate of biological productivity, and the production of calcareous material, near the surface (fig. 3.20). Beneath areas of low surface productivity, there is a small supply of sinking calcium carbonate particles and the depth of the CCD will not be substantially different from the lysocline. Beneath areas of high productivity, the large supply of sinking calcium carbonate particles will delay the point at which the supply of calcium carbonate equals the rate of dissolution. This will deepen the CCD below the lysocline. Calcareous ooze tends to accumulate on the sea floor at depths above the CCD and is generally absent at depths below the CCD. The CCD has an average depth of about 4500 m (14,800 ft), or roughly midway between the depth of the crests of ocean ridges and the deepest regions of the abyssal plains. In the Pacific, the CCD is generally at depths of about 4200–4500 m (13,800–14,800 ft). An exception to this is the deepening of the CCD to about 5000 m (16,400 ft) in the equatorial Pacific, where high rates of biological productivity result in a large supply of calcareous material. In the North Atlantic and parts of the South Atlantic, it is at or just below depths of 5000 m (16,400 ft). Calcareous oozes are found at temperate and tropical latitudes in shallower areas of the sea floor such as the Caribbean Sea, on elevated ridge systems, and in coastal regions.

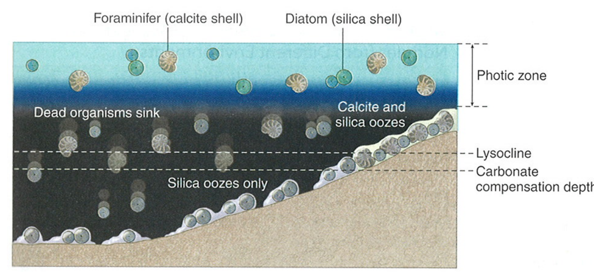


Figure 3.19 The depth at which calcium carbonate first begins to dissolve is called the lysocline. Below the lysocline, less and less calcium carbonate material is preserved. The carbonate compensation depth is the depth at which the rate of supply of calcium carbonate equals the rate of dissolution.

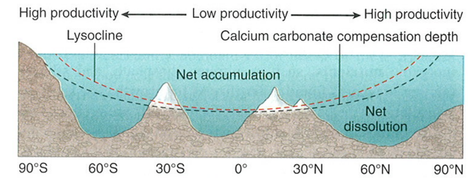


Figure 3.20 The actual depths of the lysocline and carbonate compensation depth vary regionally. The depths may be roughly equal beneath areas of low surface biological productivity where relatively little calcareous skeletal material is produced. Beneath areas of high biological productivity, the amount of calcareous material added to the water can be high, depressing the depth of the CCD below the lysocline.

Siliceous tests are created by another group of phytoplankton called diatoms and a type of zooplankton called radiolaria (fig. 3.18*a* and *b*). Their skeletal remains are the dominant components of diatomaceous and radiolarian ooze, respectively. The pattern of dissolution of siliceous tests is opposite to that of calcareous tests. The oceans are undersaturated in silica everywhere, so siliceous material will dissolve at all depths, but it dissolves most rapidly in shallow, warm water. Siliceous oozes are only preserved below areas of very high biological productivity in the surface waters (see fig. 3.17). Even in these areas, an estimated 90% or more of the siliceous tests are dissolved, either in the water or on the sea floor.

Diatomaceous ooze is found at cold and temperate latitudes around Antarctica and in a band across the North Pacific. Because diatoms are photosynthetic, they require sunlight and inorganic **nutrients,** such as nitrate, phosphate, and silicate (similar to what is found in fertilizers) for growth. The sunlight is available at the ocean's surface; the nutrients are produced by the decomposition of organisms in the ocean, and these nutrients are liberated in the deeper water as decomposition takes place. Only at certain locations are these nutrients returned to the surface by the large-scale upward flow of deeper water. Where this upward flow occurs, sunlight combines with the nutrients to produce the conditions needed for high levels of phytoplankton production. Large numbers of diatoms are found in the areas that combine suitable light, nutrients, and the correct temperature.

Radiolarian ooze is found beneath the warm waters of equatorial latitudes. Radiolaria thrive in warm water, producing siliceous outer shells that are often covered with long spines. Figure 3.18*b* shows radiolaria tests at high magnification.

Sediments derived from the water are classified as **hydrogenous** (*hydro* = water, *generare* = to produce) **sediments.** Hydrogenous sediments are produced in the water by chemical reactions. Most are formed by the slow precipitation of minerals onto the sea floor, but some are created by the precipitation of minerals in the water column in plumes of recirculated water at hydrothermal vents along the ocean ridge system. Hydrogenous sediments include some **carbonates** (limestone-type deposits), **phosphorites** (phosphorus in the form of phosphate in crusts and nodules), **salts,** and **manganese nodules.** In addition, hydrothermally generated sulfides rich in iron and other metals form along the axis of spreading centers on young sea floor, and carbonates and magnesium-rich minerals form off the axis of spreading centers on older sea floor, as discussed in chapter 2.

Hydrogenous carbonates are known to form by direct precipitation in some shallow, warm-water environments as a result of an increase in water temperature or a slight decrease in the acidity of the water. In shallow, warm water with high biological productivity, photosynthetic organisms can remove enough dissolved carbon dioxide in the water to decrease the acidity of the water and trigger the precipitation of calcium carbonate (see the discussion of carbon dioxide as a buffer in chapter 5). The calcium carbonate often precipitates in small pellets called **ooliths** (*oon* = egg) about 0.5–1.0 mm (0.02–0.04 in) in diameter. In the present oceans, there are relatively few places where this is known to be occurring. The largest modern deposits of hydrogenous carbonates are currently forming on the Bahama Banks. Additional deposits are forming on Australia's Great Barrier Reef and in the Persian Gulf.

Phosphorites contain phosphorus in the form of phosphate and are most abundant on the continental shelf and upper part of the continental slope. They are occasionally found as nodules as much as 25 cm (10 in) in diameter or in beds of sand-size grains, but more often they form thick crusts. Most phosphate deposits on continental margins do not appear to be actively accumulating. Phosphorite deposits are currently forming in regions of high biological productivity off the coasts of southwestern Africa and Peru.

Salt deposits occur when a high rate of evaporation removes most of the water and leaves a very salty brine in shallow areas. Chemical reactions occur in the brine, and salts are precipitated or separated from solution and then deposited on the bottom. In such processes, carbonate salts are formed first, followed by sulfate salts, and then chlorides, including sodium chloride. Studies of precipitated material on the floor of the Mediterranean Sea have provided clues to its past isolation from the Atlantic Ocean.

Manganese nodules are composed primarily of manganese and iron oxides but also contain significant amounts of copper, cobalt, and nickel. They were first recovered from the ocean floor in 1873 during the *Challenger* expedition. They are found in a variety of marine environments, including the abyssal sea floor, on seamounts, along active ridges, and on continental margins. Their chemistry is related to the ocean basin they are found in as well as the specific marine environment where they have grown (tables 3.3 and 3.4). Nodules from the Pacific Ocean tend to have the highest concentrations of metals, with the exception of iron. Nodules in the Atlantic Ocean generally have the highest iron concentration. The average weight percent of manganese and iron in nodules is about 18% and 17%, respectively, while the average weight percent of nickel, cobalt, and copper varies from about 0.5% down to 0.2%. Nodules that form on continental margins are very distinct chemically. They have very high manganese concentrations combined with very low iron concentrations. The chemistry of nodules can also be influenced by their position on the sea floor with respect to other sediments. Manganese nodules may lie on top of the other sediment (see fig. 3.16*a*) or be buried at shallow depth in the sediment. Nodules lying on top of the sediment react chemically with the seawater and can become enriched in iron and cobalt. Those that are buried react with both the seawater and the sediment and can become enriched in manganese and copper. The concentric layers in a nodule typically have slightly different chemistries (see fig. 3.16*b*). This chemical layering is the result of changes in the chemistry of the seawater as the nodule grew.

On the deep-sea floor, manganese nodules form black or brown rounded masses typically 1–10 cm (0.5–4 in) in diameter, roughly the size of a golf ball or a little larger. Continental margin manganese and iron oxide deposits can take a variety of forms, from nodules similar to those found on the deep-sea floor to extensive slabs, or crusts. Most manganese nodules grow very slowly: 1–10 mm (0.004–0.04 in) per million years for deep-sea nodules, roughly 1000 times slower than accumulation rates of other pelagic sediments. Nodules grow layer upon layer, often around a hard skeletal piece such as a shark's tooth, rock fragment, or fish bone that acts as a seed, much as a pearl grows around a grain of sand. They generally form in areas of very little sediment supply from other sources or where rapid bottom currents prevent them from being deeply buried. Manganese nodules on continental margins are unique in their rapid growth, having growth rates on the order of 0.01–1 mm per year—from 1000 to 1 million times faster than their deep-sea counterparts. Manganese nodules have been mapped in all oceans except the Arctic. They are most abundant in the central Pacific north and south of the biogenous oozes along the equator (see fig. 3.17). In the Atlantic and Indian Oceans, there are higher rates of lithogenous and biogenous sedimentation and consequently fewer deposits of manganese nodules.

Table 3.3 Average Chemistry of Manganese Nodules from the Three Ocean Basins

Element	Atlantic	Pacific	Indian	Average for All Three Oceans
Mn	16.18	19.75	18.03	17.99
Fe	21.2	14.29	16.25	17.25
Ni	0.297	0.722	0.510	0.509
Co	0.309	0.381	0.279	0.323
Cu	0.109	0.366	0.223	0.233

Note: The average abundances of manganese (Mn), iron (Fe), nickel (Ni), cobalt (Co), and copper (Cu) in manganese nodules from the Atlantic, Pacific, and Indian Ocean Basins. Numbers are weight % of each metal.

Table 3.4 Average Chemistry of Manganese Nodules from Different Environments

Element	Seamounts	Active Ridges	Continental	Abyssal Depths
Mn	14.62	15.51	38.69	17.99
Fe	15.81	19.15	1.34	17.25
Ni	0.351	0.306	0.121	0.509
Co	1.15	0.400	0.011	0.323
Cu	0.058	0.081	0.082	0.233
Mn/Fe	0.92	0.81	28.9	1.04

Note: The average abundances of manganese (Mn), iron (Fe), nickel (Ni), cobalt (Co), and copper (Cu), and the manganese-to-iron ratio in manganese nodules, from different environments. Numbers are weight % of each metal.

Sediments derived from space are classified as **cosmogenous** (*cosmos* = universe, *generare* = to produce) **sediments.** Particles from space constantly bombard Earth. Most of these particles burn up as they pass through the atmosphere, but roughly 10% of the material reaches the surface of Earth. Cosmogenous particles are generally small, and those that survive the passage through the atmosphere and fall in the ocean stay in suspension in the water long enough usually to dissolve before they reach the sea floor. These iron-rich sediments are found in small amounts in all oceans, mixed in with the other sediments. The pattern of related cosmic materials can indicate the direction of the particle shower that supplied them. The particles become very hot as they pass through Earth's atmosphere and partially melt; this melting gives the particles a characteristic rounded or teardrop shape. Cosmic bodies can disintegrate and melt surface materials as they strike Earth. Their impact can cause a splash of melted particles that spray outward and produce splash-form **tektites** (fig. 3.21). Microtektites are found on the ocean floor and on land.

A brief summary of the major sediment types is given in table 3.5.

Figure 3.21 Examples of splash-form tektites. These splash-form tektites were collected in Southeast Asia.

Table 3.5 Sediment Summary

Type	Source	Areas of Significant Deposit	Examples	Percent of Ocean Floor Area Covered
Lithogenous (terrigenous)	Eroded rock, volcanoes, airborne dust	Dominantly neritic, pelagic in areas of low productivity	Coarse beach and shelf deposits, turbidites, red clay	~45%
Biogenous	Living organisms	Regions of high surface productivity, areas of upwelling, dominantly pelagic, some beaches, shallow warm water	Calcareous ooze (above the CCD), siliceous ooze (below the CCD), coral	~55%
Hydrogenous	Chemical precipitation from seawater	Mid-ocean ridges, areas starved of other sediment types, neritic and pelagic	Metal sulfides, manganese nodules, phosphates, some carbonates	~1%
Cosmogenous	Space	Everywhere but in very low concentration	Meteorites, space dust	Trace

Patterns of Deposit on the Sea Floor

The patterns formed by the sediments on the sea floor reflect both distance from their source and processes that control the rates at which they are produced, transported, and deposited. Seventy-five percent of marine sediments are terrigenous. The majority of terrigenous sediments are initially deposited on the continental margins but are moved seaward by the waves, currents, and turbidity flows that move across the continental shelves and down the continental slopes. The terrigenous sediments of coastal regions are primarily lithogenous, supplied by rivers and wave erosion along the coasts. Worldwide river sediment transport is about 12–15 $\times$ 10^9 metric tons per year. The majority of this sediment enters the tropical and subtropical oceans.

Coarse sediments are concentrated close to their sources in high-energy environments—for example, beaches with swift currents and breaking waves. The waves and currents move quite large rock particles in the shore zone, but these larger particles settle out quickly. Finer particles are held in suspension and are carried farther away from their source. This pattern results in a gradation by particle size: coarse particles close to shore and to their source, with finer and finer particles predominating as the distance from the source increases.

Finer sediments are deposited in low-energy environments, offshore away from the currents and waves or in quiet bays and estuaries. In higher latitudes, deposits of rock and gravel carried along by glaciers are found in coastal environments, whereas in low latitudes, fine sediments predominate and are considered to be products of large rivers, heavy rainfall, and loose surface soils.

At the present time, most of the land-derived sediments are accumulating off the world's river mouths and in estuaries. Estuaries and river deltas serve as sediment traps, preventing terrigenous sediments from reaching the deep-sea floor in such places as the Chesapeake and Delaware Bay systems along the North Atlantic coast, in the Georgia Strait of British Columbia, and in California's San Francisco Bay along the North Pacific coast. If sediments are supplied to a delta faster than they can be retained, the sediments will move across the shelf into the deeper water environments. This is currently the case with the sediments of the Mississippi River. Much of the thick sediment layer on the outer continental shelf was laid down during the ice ages, when the sea level was lower—for example, at Georges Bank southeast of Cape Cod. Little is currently being added to these outer regions of the continental shelf.

The accumulation of sediments on the passive shelves of continents results in unstable, steep-sided deposits that may slump, sending a flow of terrigenous sediment moving rapidly down the continental slope in a turbidity current. Turbidity currents move coarse terrigenous materials farther out to sea; in doing so, they distort the general deep-sea sediment pattern and reduce the abundance of pelagic deposits. Near shore, the spring flooding of rivers alternates with periods of low river discharge in summer and fall. The floods bring large quantities of sediment to the coastal waters, and the contributions of this flooding are recorded in the layering of the sediments. Sudden masses of sediment from the collapse of a cliff or the eruption of a volcano are seen in the sediment pattern as specific additions of large quantities of sand or ash.

Figure 3.22 A deep-sea sediment core obtained by the drilling ship *Glomar Challenger*. Note the layering of the sediments.

Oceanic sediments form visually distinct layers characterized by color, particle size, type of particle, and supply rate (fig. 3.22). Seasonal variations and the patterns of long and short growing seasons for marine life can also be determined from the properties and thicknesses of the layers of biogenous material. Over long periods of geologic time, climatic changes such as the ice ages have altered the biological populations that produce sediment and have left a record in the sediment layers.

In shallow coastal areas, cycles of climate change cause variation in rates of sediment production. Along passive continental margins, biogenous sediments may also be diluted by large amounts of lithogenous sediment washing from the land. In coastal areas where marine life is very abundant and river deposits are sparse, biogenous sediments are formed from both shell fragments and broken corals. In the more homogeneous environment of the deep sea, biogenous sediments make up the majority of the pelagic deposits. There is less dilution with terrigenous materials, and few environmental changes disturb the deep bottom deposits, allowing them to remain relatively unchanged for long periods of time. Calcareous oozes are found where the production of organisms is high, dilution by other sediments is small, and depths are less than 4000 m (13,000 ft). (See the areas including the mid-ocean ridges and the warmer shallower areas of the South Pacific in fig. 3.17.) Siliceous oozes cover the deep-sea floor beneath the colder surface waters of 50°–60°N and S latitude and in equatorial regions where cold, deeper water is brought to the surface by vertical circulation. Deep basin areas of the Pacific have extensive deposits of red clay (again, see fig. 3.17).

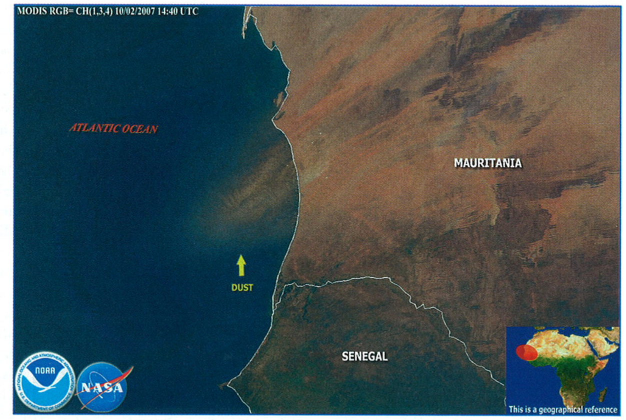


Figure 3.23 Satellite image taken October 2, 2007, showing wind-blown dust from the western Sahara Desert moving over the Atlantic Ocean.

Large rock particles of land origin are also moved out to sea by a process known as **rafting.** Glaciers carry sand, gravel, and rocks embedded in the ice. When the glacier reaches the sea, parts break off and fall into the water as icebergs. The icebergs are carried away from land by the currents and winds, taking the terrigenous materials far from their original sources. As the ice melts, rocks and gravel that were frozen in the ice sink to the sea floor. In addition, sea ice formed in shallow water along the shore can incorporate material from the sea floor and transport it out to sea. Figure 3.17 indicates areas of terrigenous deposits that are affected by ice rafting. It is estimated that ice-rafted material can be found over about 20% of the sea floor. Sometimes large, brown seaweeds known as kelp, which grow attached to rocks in coastal areas, are dislodged by storm waves. The kelp may have enough buoyancy to float away, carrying the attached rock. When the seaweed dies or sinks, the rock is deposited on the ocean floor at some distance from its origin. The deposition of larger rocks by this rafting process is infrequent and irregular.

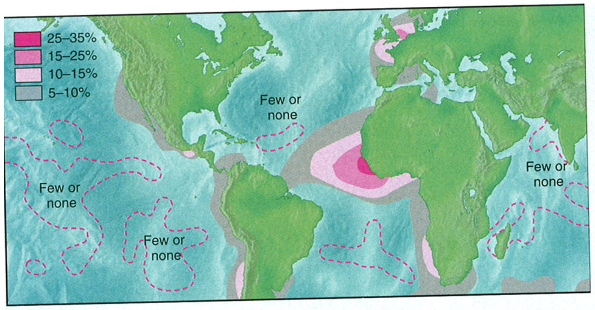
The wind is an effective agent for moving lithogenous materials out to sea in some parts of the world. Winds blowing offshore from the Sahara Desert or other arid regions transfer sand particles directly from land to sea, sometimes 1000 km (600 mi) or more offshore (fig. 3.23). A similar process can occur between sand dunes and coastal waters. In the open ocean, airborne dust probably supplies much of the deep-sea red clay material. Figure 3.24 indicates the frequency with which winds carry dust, or haze, out to sea. The world's volcanoes are another source of airborne particles. Volcanic ash is present in seafloor sediments and can be found in layers of significant thickness associated with past volcanic events.

Formation of Rock

Loose sediments on the sea floor are transformed into **sedimentary rock** in a process known as **lithification.** Lithification can occur through burial, compaction, recrystallization, and cementation. As one layer of sediment covers another, the weight of the sediments puts pressure on the lower sediment layers, and the sediment particles are squeezed more and more tightly together. The particles begin to stick to each other, and the pore water between the sediment particles, with its dissolved solids, moves through the sediments. As it does so, minerals precipitate on the surfaces of the particles and, in time, act to cement the sediment particles together into a mass of sedimentary rock. The sediments in these processes are also exposed to increasing temperature with increasing depth of burial. Chemical changes also occur in sedimentary particles through interaction with seawater and pore water in a process called **diagenesis.** One example of diagenesis is the gradual lithification of calcareous ooze to form chalk or limestone. In this process, calcite particles in the sediments are cemented by calcite precipitated from the pore waters. The transformation of calcareous ooze to chalk occurs at a sediment depth of a few hundred meters, and the further transformation to limestone occurs with additional cementation under about 1 km burial. Siliceous oozes can be lithified to form a very hard rock called chert.

Sedimentary rock may preserve the layering of the sediments in visually distinct features and strata. Ripple marks from the motion of waves and currents may be seen, and fossils may also be present. Sedimentary rocks are found beneath the sediments of the deep-sea floor, along the passive margins of continents, and on land where they have been thrust upward along active margins or formed in ancient inland seas. Sedimentary rocks include sandstone, shale, and limestone.

If sediments are subjected to greater changes in temperature, pressure, and chemistry, **metamorphic rock** results. Slate is a metamorphic rock derived from shale, and marble is recrystallized metamorphosed limestone.



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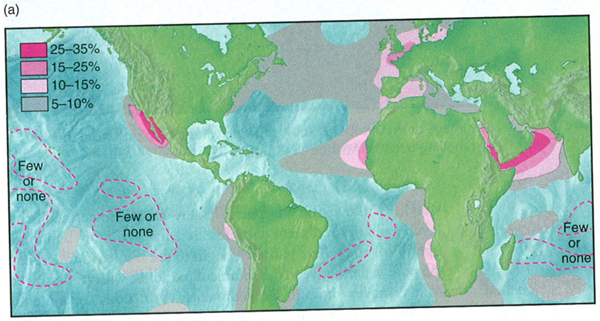


(b)

Figure 3.24 Frequency of haze as a result of airborne dust during the Northern Hemisphere's (a) winter and (b) summer. Values are given in percentages of total observations.

Sampling Methods

To analyze sediments, the geological oceanographer must have an actual bottom sample to examine. A variety of devices have been developed to take a sample from the sea floor and return it to the laboratory for analysis. **Dredges** are net or wire baskets that are dragged across a bottom to collect loose bulk material, surface rocks, and shells in a somewhat haphazard manner (fig. 3.25). **Grab samplers** are hinged devices that are spring- or weight-loaded to snap shut when the sampler strikes the bottom. See figure 3.26 for examples of this device. Grab samplers sample surface sediments from a fixed area of the sea floor at a single known location.

A **corer** is essentially a hollow pipe with a sharp cutting end. The free-falling pipe is forced down into the sediments by its weight or, for longer cores, by a piston device that enables water pressure to help drive the core barrel into the sediment; coring devices are shown in figure 3.27*a–e*. The product is a cylinder of mud, usually 1–20 m (3.3–65.6 ft) long, that contains undisturbed sediment layers (see fig. 3.22). Box corers (fig. 3.27*e*) are used when a large and nearly undisturbed sample of surface sediment is needed. These corers drive a rectangular metal box into the sediment; they have doors that close over the bottom before the sample is retrieved. Long cores that penetrate the thick sediment overlying older sea floor and reach the older sediment layers nearer the oceanic basalt may be obtained by drilling through both loose sediments and rock. The highly sophisticated drilling techniques used by the research vessel *JOIDES Resolution* are discussed in chapter 2.

Geologic oceanographers and geophysicists also study sediment distribution and seafloor structure with high-intensity sound, a technique known as **acoustic profiling.** Bursts of sound are directed toward the sea floor, where the sound waves either reflect from or penetrate into the sediments. Sound waves that penetrate the sediments are refracted and change speed as they pass through the different layers of sediments. A surface vessel tows an array of underwater microphones, or hydrophones, to sense the returning sound waves, and a recorder plots the returning sound energy to produce a profile of the sediment structure. This technique details the structure of the continental margin and finds buried faults, filled submarine canyons, and clues to oil and gas deposits.

Today's ocean scientists are searching for records of Earth's history in the sediment and rock layers of the ocean floor. These layers hold evidence for understanding the formation of the ocean basins and continents, changing climate, periods of unusual volcanism, the presence and absence of various life-forms, and much more. The information is there, but it requires a combination of sophisticated technical know-how at sea and increasingly detailed scientific research in the laboratory to discern and understand it.

Sediments as Historical Records

Marine sediments and the skeletal materials in them provide important information about processes that have shaped the planet and its ocean basins over the past 200 million years. The study of the oceans through an analysis of sediments is called **paleoceanography.** Two examples of the use of marine sediments to unravel history are (1) the study of the distribution of skeletal remains of marine organisms to date the initiation of the Antarctic Circumpolar Current (ACC) and (2) the study of the relative abundance of different oxygen isotopes in foraminifera tests preserved in the sediment to determine variations in climate and seawater temperature.

Prevailing westerly winds at high southern latitudes cause the ACC to flow continuously from west to east around Antarctica. The ACC is a very deep current, extending to depths of 3000–4000 m (9800–13,000 ft), and it is able to flow unimpeded around the globe because there are no shallow seafloor features to block its path. This situation has not always existed, however. The southern continents began to break apart at different times. About 135 million years ago, Africa and India first began to separate from Antarctica, South America, and Australia (see fig. 2.39). As recently as 80 million years ago, South America, Antarctica, and Australia were still effectively one landmass. Sometime around 55 million years ago, some sea floor existed between Australia and Antarctica, and by 35 million years ago, they had separated sufficiently to create a narrow expanse of water called the Austral Gulf. South America had not yet separated from Antarctica. Marine sediments deposited at this time indicate that a small, single-celled, shallow-water marine foraminiferan called *Guembelitria* lived in the restricted waters of the Austral Gulf. The absence of its remains in other Southern Hemisphere sediments of the same age indicates that the organism had not been spread to other areas by ocean currents. Skeletal remains of *Guembelitria* appear quite suddenly in sediments deposited all around Antarctica about 30 million years ago. Even though the Drake Passage between South America and Antarctica did not fully open before 20 million years ago, there must have been a shallow channel a few hundred meters deep as early as 30 million years ago that allowed the ACC to first flow around the continent, carrying *Guembelitria* with it.

Calcite tests found in successive layers of sediment can provide information about changes in climate and seawater temperature over time through a careful analysis of the relative abundance of different oxygen isotopes in the calcite. **Isotopes** are atoms of the same element that have different numbers of neutrons in the nucleus; thus, they have different atomic masses but behave identically chemically. Some marine organisms remove oxygen from water molecules in the ocean to construct calcareous hard parts. Water contains the two main isotopes of oxygen: the common ^{16}O and the rarer ^{18}O. These isotopes are stable and do not decay radioactively, so once they have been incorporated into an organism's skeletal material, their relative proportion (^{18}O:^{16}O) remains constant even after the organism dies. The ^{18}O:^{16}O ratio in a skeletal fragment depends in part on the relative abundance of the isotopes in the seawater at the time the organism formed it. Thus, calcareous biogenous remains record changes in the isotopic chemistry of seawater that are related to changes in global temperature.

(a)

(b)

Figure 3.25 (a) Rocks can be recovered from the sea floor with a dredge having a chain basket. Sediments and other fine material escape through the chains. (b) Basalt dredged from a depth of about 8 km (5 mi) near the Tonga Trench in the western Pacific Ocean. The dredge is in the *foreground*.

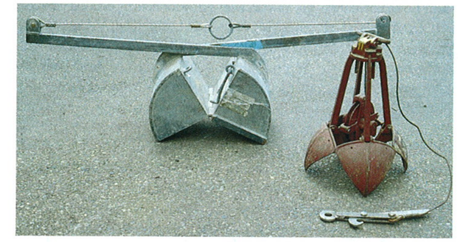
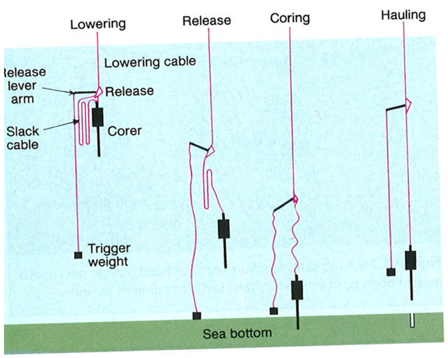


Figure 3.26 Grab samplers: Van Veen (*left*) and orange peel (*right*), both in open positions. Grabs take surface sediment samples.



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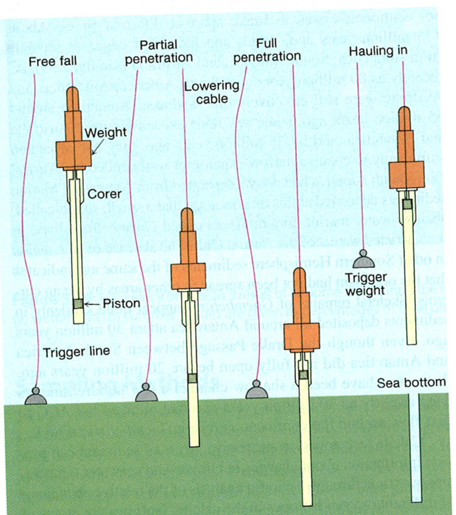

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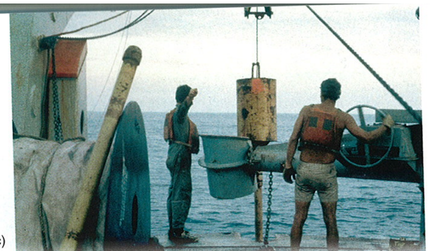

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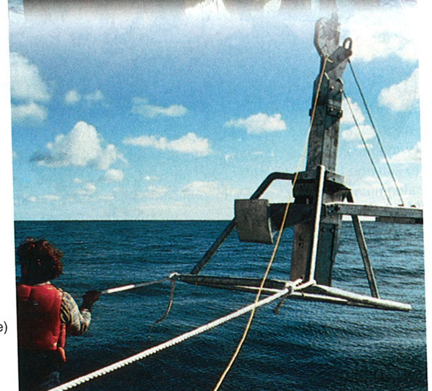


Figure 3.27 (a) The Phleger corer is a free-fall gravity corer. The weights help to drive the core barrel into the soft sediments. Inside the corer is a plastic liner. The sediment core is removed from the corer by removing the plastic tube, which is capped to form a storage container for the core. (b) A sketch of a piston corer in operation. The corer is allowed to fall freely to the sea bottom. The action of the piston moving up the core barrel owing to the tension on the cable allows water pressure to force the core barrel into the sediments. (c) Recovering a piston corer. The barrel of a gravity corer is in the left foreground. (d) A gravity corer ready to be lowered. (e) A box corer is used to obtain large, undisturbed seafloor surface samples.

Water molecules containing ^{16}O are lighter than molecules containing ^{18}O, so they are more easily removed from the oceans by evaporation. During glacial periods, the water evaporated from the sea surface is trapped in ice sheets; the sea level is lowered, and ^{16}O is removed from the ocean system. This process increases the $^{18}O{:}^{16}O$ ratio in the seawater and in skeletal parts that organisms are forming at that time. When these organisms die, their skeletal parts sink to the sea floor and are incorporated into the sediment. During warmer, interglacial periods, the melting of ice sheets causes a rise in sea level and returns ^{16}O-enriched fresh water to the oceans. The result is a drop in the $^{18}O{:}^{16}O$ ratio in the seawater and in the skeletal parts that are being formed. The isotopic composition of skeletal parts is also influenced by seawater temperature. As temperature decreases, organisms preferentially take up more ^{18}O than ^{16}O in their skeletons. The actual $^{18}O{:}^{16}O$ ratio preserved in a skeletal fragment is primarily due to changes in seawater composition related to the growth and decay of global ice sheets and consequent fall and rise of sea level.

QUICK REVIEW

1. Describe three different ways to classify marine sediments.
2. Relate the distribution of calcareous ooze to seafloor features.
3. Explain the factors that contribute to the accumulation of siliceous oozes and red clays.
4. What is the difference between the lysocline and the CCD?
5. How can small particles sink quickly to the sea floor?

3.4 Seabed Resources

Long ago, people began to exploit the materials of the seabed. The ancient Greeks extended their lead and zinc mines under the sea, medieval Scottish miners followed seams of coal under the Firth of Forth, and, more recently, coal has been mined from undersea strata off Japan, Turkey, and Canada. As technology has developed and as people have become concerned about the depletion of onshore mineral reserves, interest in seabed minerals and mining has grown. At present, the United States is showing little interest in new seabed resources, but international interest remains strong; research continues in exploration, technology development, and environmental studies, especially in Japan, India, China, and South Korea. Keep in mind that each potential deep-sea source is in competition with an onshore supply. Whether the seabed source will be developed depends largely on international markets, needs for strategic materials, and whether offshore production costs can compete with onshore costs.

Sand and Gravel

The largest superficial seafloor mining operation is for sand and gravel, widely used in construction. The technology and cost required to mine sand and gravel in shallow water differ very little from land operations. This is a high-bulk, low-cost material tied to the economics of transport and the distance to market.

Sand and gravel mining is the only significant seabed mining done by the United States at this time. It is estimated that the United States has a reserve of 450 billion tons of sand off its northeastern coast; there are large deposits of gravel along Georges Bank off New England and in the area off New York City. Along the coasts of Louisiana, Texas, and Florida, shell deposits are mined for use in the lime and cement industries, as a source of calcium oxide used to remove magnesium from seawater as part of the process of making magnesium metal, and, when crushed, as a gravel substitute for roads and highways.

Sands are mined as a source of calcium carbonate throughout the Bahamas, which have an estimated reserve of 100 billion metric tons. Coral sands are mined in Fiji, in Hawaii, and along the U.S. Gulf Coast. Other coastal sands contain iron, tin, uranium, platinum, gold, and diamonds. The "tin belt" stretches for 3000 km (1800 mi) from northern Thailand and western Malaysia to Indonesia. Here, sediments rich in tin have been dredged for hundreds of years and supply more than 1% of the world's market. Iron-rich sediments are dredged in Japan, where the reserve of iron in shallow coastal waters is estimated at 36 million tons. The United States, Australia, and South Africa recover platinum from some sands, and gold is found in river delta sediments along Alaska, Oregon, Chile, South Africa, and Australia. Diamonds, like gold, are found in sediments washed down the rivers in some areas of Africa and Australia. Muds bearing copper, zinc, lead, and silver also occur on the continental slopes, but they lie too deep for exploitation, considering the present demand and their market value.

Phosphorite

Phosphorite, which can be mined to produce phosphate fertilizers, is found in shallow waters as phosphorite muds and sands containing 12%–18% phosphate and as nodules on the continental shelf and slope. The nodules contain about 30% phosphate, and large deposits are known to exist off Florida, California, Mexico, Peru, Australia, Japan, and northwestern and southern Africa. Recently, a substantial source of phosphorite was located in Onslow Bay, North Carolina. Eight beds have been found, and five are thought to be economically valuable; they have been estimated to contain 3 billion metric tons of phosphate concentrates.

The world's ocean reserve of phosphorite is estimated at about 50 billion tons. Readily available land reserves are not in short supply, but most of the world's land reserves are controlled by relatively few nations. Therefore, political considerations may make these marine deposits attractive as mining ventures for some countries.

Oil and Gas

Oil and gas represent more than 95% of the value of all resources extracted from the sea floor or below. Oil and gas deposits are almost always associated with marine sedimentary rocks and are believed to be produced by the slow conversion of marine plant and animal organic matter to hydrocarbons. Conditions must be just right for marine organic material to eventually be converted to oil and gas. It must first accumulate in relatively shallow, quiet water with low oxygen content. Anaerobic bacteria can then utilize the organic matter to produce methane and other light hydrocarbons. As these simple hydrocarbons are buried beneath deeper layers of sediment, they are subjected to higher pressure and temperature. Over a period of millions of years, they can be converted to oil or gas. Oil forms if the depth of burial is on the order of about 2 km (1.2 mi). If the organic material is buried even deeper or cooked for a longer period of time at higher temperature, gas is produced. Oil deposits are generally found at depths less than 3 km (1.9 mi), and below 7 km (4.3 mi) only gas is found.

Because oil and gas are very light, they migrate upward over time, moving slowly out of the source rock and into porous rocks above. This upward migration continues until the fluids reach an impermeable layer of rock. The oil and gas then stop their ascent and fill the pore spaces of the reservoir rock below this impermeable layer.

Petroleum-rich marine sediments are more likely to accumulate during periods of geologic time when sea level is unusually high and the oceans flood extensive low-lying continental regions to create large shallow basins. Much oil and gas are found in marine rocks that formed from sediments deposited during a relatively short period of time during the Jurassic and Cretaceous, between about 85 million and 180 million years ago, when sea level was high.

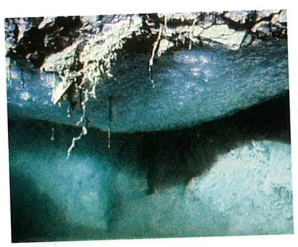
Major offshore oil fields are found in the Gulf of Mexico, the Persian Gulf, and the North Sea, and off the northern coast of Australia, the southern coast of California, and the coasts of the Arctic Ocean.

Bringing the offshore oil fields into production has required the development of massive drilling platforms and specialized equipment to withstand heavy seas and fierce storms and to allow drilling and well development at great depth. Although the cost of drilling and equipping an offshore well is three to four times greater than that of a similar venture on land, the large size of the deposits allows offshore ventures to compete successfully. The gas and oil potential in even deeper offshore waters is still unknown, but the deeper the water in which the drilling must be done, the higher the cost.

The new methods and equipment developed and used for deep-sea oceanographic drilling and research have provided the prototypes for new generations of deep-sea commercial drilling systems. Even though legal restraints, environmental concerns, and worldwide political uncertainties will continue to contribute to the slow development of offshore deposits, petroleum exploration and development will undoubtedly continue to be the main focus of ocean mining in the near future.

Gas Hydrates

In recent years, interest has been growing in gas hydrates trapped in marine sediments (fig. 3.28). Gas hydrates are a combination of natural gas, primarily methane (CH_4), and water, which forms a solid, icelike structure under pressure at low temperatures. Drill cores of marine sediment have recovered samples of gas hydrates that melt and bubble as the natural gas escapes. These melting samples burn if lit. Gas hydrates are a subject of intense interest for three reasons: they are a potential source of energy, they may contribute to slumping along continental margins, and they may play a role in climate change.



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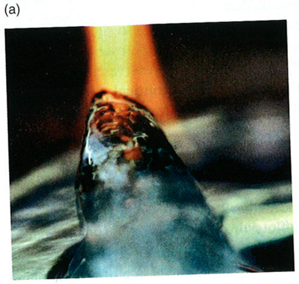

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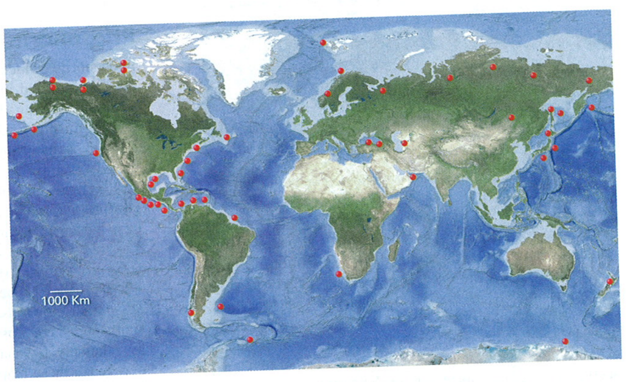

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Figure 3.28 (a) The cream-colored, icy material in this underwater ledge is gas hydrate. Frozen water can trap other molecules within its cage-like structure, including molecules of methane gas. (b) Gas hydrate is called the *ice that burns*. Here, the methane being released from the ice is burning, not the water ice, which is incombustible. (c) Gas hydrates occur within ocean-floor sediments, especially in parts of the ocean that are cold and deeper than 500 m (1,600 ft). Common settings are along passive margins and trenches. Gas hydrates also occur on land, beneath the frozen arctic tundra.

When 1 cubic foot of gas hydrate melts, it releases about 160 cubic feet of gas. A gas hydrate accumulation thus can contain a huge amount of natural gas. Estimates of the amount of natural gas contained in the world's gas hydrate accumulations are speculative and range over three orders of magnitude, from about 2800 to 8,000,000 trillion (2.8×10^{15} to 8×10^{18}) cubic meters of gas. By comparison, in 2000, the U.S. Geological Survey estimated conventional natural gas accumulations for the world at approximately 440 trillion (4.4×10^{14}) cubic meters. Despite the enormous range in the estimated amount of natural gas contained in gas hydrates, even the lowest estimates suggest that gas hydrates are a much greater resource of natural gas than conventional accumulations and may be a substantial source of energy in the future (table 3.6). It is important to note, however, that none of these assessments have predicted how much gas could actually be produced from the world's gas hydrate accumulations given present technology and their location.

A second reason gas hydrates are significant is their effect on seafloor stability. Along the southeastern coast of the United States, a number of submarine landslides, or slumps, have been identified that may be related to the presence of gas hydrates. The hydrates may inhibit normal sediment consolidation and cementation processes, creating a weak zone in the sediments. Alternately, the lowering of sea level during the last glacial period may have reduced the pressure on the sea floor enough to allow some of the gas to escape from the hydrates and accumulate in the sediment, decreasing its strength.

A final reason for studying gas hydrates is their potential link to climate changes. The amount of methane stored in hydrates is believed to be about 3000 times the amount currently present in the atmosphere. Since methane is a greenhouse gas, its release from hydrates could affect global climate.

Manganese Nodules

Manganese nodules are found scattered across the world's deep-ocean floors, with particular concentrations in the red clay regions of the northeastern Pacific (see figs. 3.16 and 3.17). The nodule chemistry varies from place to place, but the nodules in some areas contain 30% manganese, 1% copper, 1.25% nickel, and 0.25% cobalt: these are much higher concentrations than are usually found in land ores. Cobalt is of particular interest, since it is classified as being of "strategic" importance to the United States and hence essential to the national security. Cobalt is an important component in the manufacture of strong alloys used in tools and aircraft engines. The nodules grow very slowly, but they are present in huge quantities. An estimated 16 million additional tons of nodules accumulate each year.

Cobalt-enriched manganese crusts, or hard coatings on other rocks, were discovered in relatively shallow water on the slopes of seamounts and islands within U.S. territorial waters in the 1980s. The concentration of cobalt in these deposits is roughly twice that found in typical pelagic manganese nodules and about one and one-half times that found in known continental deposits. These crusts are not being actively mined because of the relatively low cost and continued availability of continental sources.

Sulfide Mineral Deposits

Expeditions to the rift valleys of the East Pacific Rise near the Gulf of California, the Galápagos Ridge off Ecuador, and the Juan de Fuca and Gorda Ridges off the northwestern United States have found sulfides of zinc, iron, copper, and possibly silver, molybdenum, lead, chromium, gold, and platinum. Molten material from beneath Earth's crust rises along the rift valleys, fracturing and heating the rock. Seawater percolates into and through the fractured rock, forming metal-rich hot solutions. When these solutions rise from the cracks and cool, the metallic sulfides precipitate to the sea floor. Deposits may be tens of meters thick and hundreds of meters long. Too little is presently known about these deposits to determine whether they might be of economic importance at some future date. No practical technology exists to sample or retrieve them at this time, and, like the manganese nodules, these deposits are found outside national economic zones, so there are ownership problems.

In the 1960s, metallic sulfide muds were discovered in the Red Sea. Deposits of mud 100 m (330 ft) thick were found in small basins at depths of 1900–2200 m (6200–7200 ft). High amounts of iron, zinc, and copper and smaller amounts of silver and gold were found. The salty brines over these muds contained hundreds of times more of these metals than normal seawater.

Table 3.6 Potential Significance of Gas Hydrates

Estimated Volume of Gas Hydrates, EVGH (10^{12} m^3)	Ratio of EVGH to World Supply of Natural Gas	Ratio of EVGH to Natural Gas Consumption in the: United States in 2000	World in 1999
Low of 2800	6.4:1	4375:1	1175:1
High of 8,000,000	18,200:1	12,500,000:1	3,355,700:1

QUICK REVIEW

1. How are sand and gravel used as valuable resources both directly and indirectly?
2. What seafloor resource is used in the production of fertilizer?
3. What resources account for the vast majority of the monetary value of all resources extracted from the ocean?
4. List multiple reasons why gas hydrates are important.
5. What important metals are found in manganese nodules?
6. Hydrothermal vents along mid-ocean ridges are often the site of what natural resource?

Summary

Ocean-depth measurements were made first with a hand line, then with wire, and, since the 1920s, with echo sounders. Today, they are made with precision depth recorders. Seafloor features can also be sensed by satellites that measure the distance between the satellite and the sea surface.

The bathymetric features of the ocean floor are as rugged as the topographic features of the land but erode more slowly. The continental margin includes the continental shelf, slope, and rise. The continental shelf break is located at the change in steepness between the continental shelf and the continental slope. Submarine canyons are major features of the continental slope and, in some cases, the continental shelf. Some canyons are associated with rivers; others are believed to have been cut by turbidity currents. Turbidity currents deposit graded sediments known as turbidites.

The ocean basin floor is a flat abyssal plain, but it is interrupted by scattered abyssal hills, volcanic seamounts, and flat-topped guyots. In warm, shallow water, corals have grown up around the seamounts to form fringing reefs. A barrier reef is formed when a seamount subsides while the coral grows. An atoll results when the seamount's peak is fully submerged. The mid-ocean ridges and rises extend through all the oceans; trenches are associated with island arcs and are found mainly in the Pacific Ocean.

Sediment classifications are based on their size, location, origin, and chemistry. Sediment particles are broadly categorized in order of decreasing size as gravel, sand, and mud. Within each of these categories, particles can be further subdivided by size. The sinking rate and distance traveled in the water column are related to sediment size, shape, and currents. Small particles sink more slowly than large particles. The very smallest particle sizes, silts and clays, sink so slowly that they may be transported large distances while falling to the sea floor. The sinking rate of particles is increased by clumping and incorporation into fecal pellets.

Sediments that accumulate on continental margins and the slopes of islands are called neritic sediments. Sediments of the deep-sea floor are pelagic sediments. In general, pelagic sediments accumulate very slowly and neritic sediments accumulate more rapidly.

Sediments formed from particles of preexisting rocks are called lithogenous sediments. These sediments are also sometimes called *terrigenous sediments.* Since lithogenous sediments are typically derived from the land, they are also known as terrigenous sediments. Pelagic lithogenous sediment is dominated by red clay. Red clay dominates marine sediments only in regions that are starved of other sources of sediment. Biogenous sediments come from living organisms. Sediments composed of at least 30% biogenous material are called oozes; this material accumulates in regions of high biological productivity. Siliceous sediments are subjected to dissolution everywhere in the oceans, while calcareous sediments dissolve rapidly in deep, cold water below the CCD. Sediments that precipitate directly from the water are called hydrogenous sediments. These include manganese nodules on the deep-sea floor and metal sulfides along mid-ocean ridges. Sediments containing particles that originate in space are called cosmogenous sediments.

Patterns of sediment deposit result from the distance from the source area, the abundance of living organisms contributing remains, the seasonal variations in river flow, waves and currents including turbidity currents, the variability in land sources, the prevailing winds, and sometimes rafting.

Coarse sediments are concentrated close to shore; finer sediments are found in quiet offshore or nearshore environments. Terrigenous sediments are found mainly along coastal margins; most deep-sea sediments come from biogenous sources. The distributions of particle sizes reveal the processes that formed the deposit, and the sediment layers provide clues to ancient climate patterns.

In general, sedimentation rates are slowest in the deep sea and greatest near the continents. In some areas, relict sediments were deposited under conditions that no longer exist. Mechanisms that increase the sinking rates of particles include clumping and incorporation of sediment particles into larger fecal pellets of small marine organisms. Loose sediments are transformed into sedimentary rock in which the layering of the sediments may be preserved.

Sediments are sampled with dredges, grabs, and corers; deep-sea drilling takes samples through the sediments and from the seafloor rock below.

Calcareous biogenous sediments preserve records of changes in the oxygen isotopic composition of seawater that are related directly to water temperature and hence can be used to study changes in global climate. Consequently, variations in ^{18}O:^{16}O isotopic ratios in calcareous skeletal remains record fluctuations in global coverage by ice sheets and in sea level.

Seabed resources include sand and gravel used in construction and landfills. Sands and muds that are rich in mineral ores are mined. Phosphorite nodules are the raw material of fertilizer. Oil and gas are the most valuable of all seabed resources. Manganese nodules are rich in copper, nickel, and cobalt; they are present on the ocean floor in huge numbers. Retrieval of seafloor mineral resources is slowed by disputes over international law, high mining costs, and low market prices. Sulfide mineral deposits have been discovered along rift valleys; their economic importance is unknown.

Large deposits of gas hydrates are being studied to determine their potential as economically important sources of methane gas. These deposits are icelike accumulations of natural gas and water that form at low temperature and high pressure on the sea floor. Scientists are also studying their possible role in submarine landslides and global climate change.

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Study Problems

1. Assume an accumulation rate of 0.8 cm per 1000 years for deep-ocean pelagic sediment. How long would it take to accumulate 500 m of sediment?
2. Assume an accumulation rate of 30 cm per 1000 years for continental shelf sediment. How long would it take to accumulate 500 m of sediment?
3. If underwater cables are spaced 14 km apart on the sea floor, and if monitoring equipment shows that they break in sequence from the shallowest to the deepest at 15-minute intervals, what can you determine about the event that caused the breaks?
4. If the average concentration of suspended sediment in the water of a harbor is 1 g/m^3, the volume of water in the harbor is 158 km^3, and the daily sediment supply rate averages 1×10^7 kg, what is the average residence time of sediment in the water? If the harbor has an average depth of 15.8 m, what is the surface area of the harbor?