Introduction to Ocean Sciences

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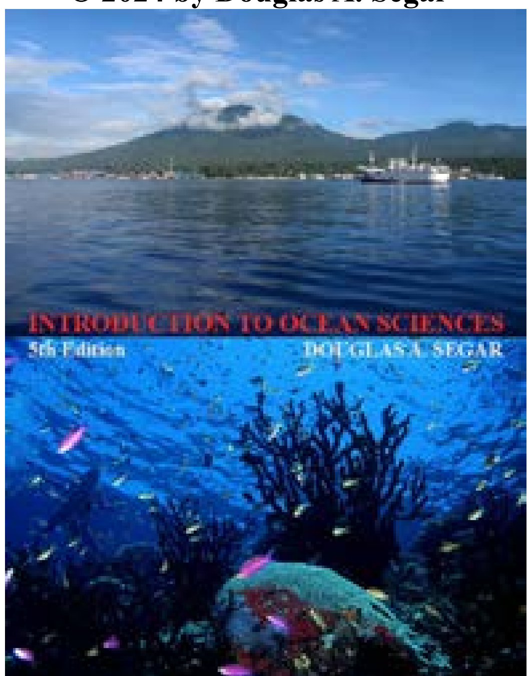

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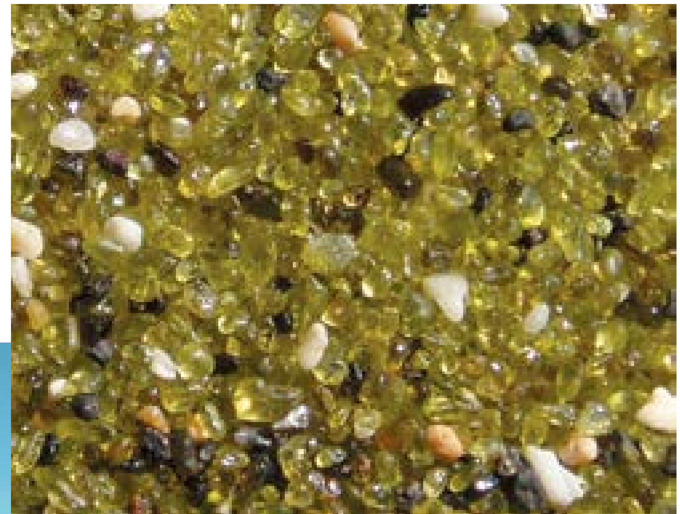
CHAPTER 6

Ocean Sediments

CRITICAL CONCEPTS USED IN THIS CHAPTER
CC1 Density and Layering in Fluids
CC2 Isostasy, Eustasy, and Sea Level
CC3 Convection and Convection Cells
CC4 Particle Size, Sinking, Deposition, and Resuspension
CC6 Salinity, Temperature, Pressure, and Water Density
CC7 Radioactivity and Age Dating
CC8 Residence Time
CC9 The Global Greenhouse Effect
CC14 Phototrophy, Light, and Nutrients



This beautiful cove and beach on the south shore of Hawaii are known locally as the "Green Sand Beach." The shoreline consists of volcanic ash that is easily eroded. The eroded cliffs supply the source of the sand on this beach and those nearby, as well as on the seafloor on the narrow continental shelf. The minerals are not weathered and have a distinctly greenish hue that, unfortunately, does not show well in photographs but that can be seen more easily in the inset close-up image of the sand grains.

Why should we be interested in marine **sediments**? Approximately 70% of the Earth's surface is covered by the oceans, and almost all of the seafloor beneath these oceans is covered by **sediments.** These sediments provide a home for many living species and contain a wealth of information about the history of the Earth's climate, oceans, and continents. Sediments accumulate layer by layer and include the remains of dead organisms. By studying changes in the type and chemical composition of the organisms, and the chemistry of sediment in each layer beneath the sediment surface, we can learn a great deal about the Earth's history and the evolution of life.

This chapter reviews the origin and types of particles present in ocean sediments and the mechanisms by which they enter the oceans and are transported to the seafloor. It also describes how the type of sediment that accumulates at any location can change with climate and plate tectonics and examines how the sediments can be used to investigate the Earth's history.

SEDIMENTS AND BIOGEOCHEMICAL CYCLES

Sediments are important components of **biogeochemical** cycles (**Chap. 5**), and they act as both sinks and sources of elements dissolved in seawater. Human civilization has now altered the **steady state** (**Chap. 5**) of the biogeochemical cycles of many elements by mining rocks and burning **fossil fuels**; by deforestation, which increases **erosion**; and by releasing **contaminants** that cause **acid rain** and promote **leaching**. Many human activities have increased the rate at which some elements enter the ocean and atmosphere, which will probably cause concentrations of these elements in seawater to rise, even if the rate of removal to the sediment increases to match the increased input rate (**CC8**).

For many elements, the rate of input to the oceans has varied over millions of years. Periods of intense volcanic activity or changes in land erosion rates related to sea-level and **glaciation** changes, for example, affect the rate of input to the oceans. Without a full understanding of these processes, we cannot understand or predict how releases of elements by human activities will affect the ocean environment. Biogeochemical cycles are difficult to study because of their many complex interactions and because some of the key processes occur on timescales of tens of millions of years. Fortunately, seafloor sediments preserve a history of the biogeochemical cycles and their variations over the past 170 million years (the age range of almost all surviving seafloor; **Chap. 4**) and perhaps as far as 340 million years ago. Sedimentary rocks on land preserve even older history. Studying the jigsaw puzzle of information contained in ocean sediment is essential to understanding the consequences of human activities, including pollution (**Chap. 16**).

CLASSIFICATION OF SEDIMENTS

Sediments are separated into categories to facilitate comparisons and communication among scientists. There are a number of classification schemes. The most widely used schemes are based either on the range of sizes of the particles (grains) that make up the sediment or on the origin of the predominant particles in the sediment. Each type of scheme has its advantages, as we shall see in this chapter.

Classification by Grain Size

The size of the grains in marine sediments varies from boulders tens of centimeters or more in diameter to grains that are so small (less than 0.001 mm in diameter) that they cannot be distinguished except under the most powerful electron microscopes. Various grain size classification schemes have been used, but generally sediments are described as clay, silt, sand, or gravel (**Fig. 6-1**). Sediments with the finest grain size ranges—silt and clay—are often classified together as mud.

Sediments with a wide range of grain sizes can be found in different areas of the oceans or at different depths within the layers of sediment at any given location. However, most marine sediments that have accumulated at the same time and in the same area of the oceans consist of only grains that are within a narrow range of grain sizes. Sediments that consist mostly of grains that are within a narrow range of sizes are said to be well sorted. For example, most sediment on the deep-ocean floor consists of well-sorted mud, and many areas near beaches consist of well-sorted sand. Sediments that consist of a wide range of grain sizes are said to be poorly sorted. For example, some undersea landslide deposits are poorly sorted, consisting of a mixture of mud, sand, and gravel. Poorly sorted sediments are found in only limited areas in the oceans. Transport mechanisms that cause sediment sorting are discussed in **CC4**, and sorting of beach sand is discussed in **Chapter 11**.

Classification by Origin

Sediments can be classified as **lithogenous**, **biogenous**, **hydrogenous**, or **cosmogenous** to indicate the origin of the dominant type of particle. Lithogenous particles consist of fragments of rocks and minerals **weathered** and eroded from rocks on land, and fragments of ash and rock from terrestrial and undersea volcanic eruptions. Biogenous particles are the remains of organisms. Because most organic matter is decomposed rapidly, biogenous particles are predominantly the mineralized **hard parts** (shells or skeletal material) of marine organisms. They include, for example, the hard parts of certain **phytoplankton** and **zooplankton** species, **mollusk** shells, fish bones, and whale and shark teeth. Hydrogenous particles are formed by precipitation of dissolved inorganic chemicals from seawater to form solids. Cosmogenous particles are meteorite fragments that have passed through the atmosphere.

Each category of particles can be subdivided into several different types. For example, biogenous particles can be classified according to the types of organisms from which they are derived (**Fig. 6-1**).

The distribution of sediment particles in the oceans is determined by the size range and origin of each type of particle, the susceptibility of each type of particle to decomposition or dissolution in seawater, and the mechanisms transporting the particles. The distribution of sediments is discussed later in this chapter. However, to understand the distribution of sediments, we need first to understand the origins and sources of the particles that make up sediments and the processes that transport and transform the particles in the ocean.

LITHOGENOUS SEDIMENTS

Most lithogenous sediment particles are the products of **weathering** and erosion of terrestrial rocks by water and wind. For this reason, sediments with a high proportion of lithogenous particles are often called **terrigenous** sediments. Rock fragments are continuously flaked off of solid rock by the action of running water or waves, as a result of freezing and thawing of ice, or by the actions of plant roots or animals. Once they are broken into fragments, rock particles are further weathered to smaller particles and transported by streams, rivers, waves, winds, and **glaciers**.

Rocks of the landmasses are composed of a large variety of minerals. The minerals are slowly altered and partially dissolved by reactions with oxygen, carbon dioxide, and water. During such chemical weathering, many minerals partly dissolve, leaving behind resistant minerals including quartz, feldspars, and clay minerals, which are all siliceous minerals. Clay minerals are layered structures of silicon, aluminum, and oxygen atoms. Some clay minerals also contain iron or other elements.

Five natural transport mechanisms bring lithogenous sediment to the oceans: freshwater runoff, glaciers, waves, winds, and landslides. Humans have added another mechanism: ships. Until the mid to late 20th century, all vessels simply threw trash and garbage overboard. Cans, bottles, plastics, and clinker (cinders from coal-burning ships) can now be found in sediments throughout the oceans, particularly under major shipping lanes (**Chap. 16**). The various transport mechanisms introduce different size ranges of sediment particles to the oceans, and each mechanism has a different distribution of input locations.

Transport by Rivers

Rivers vary substantially in terms of their volume of flow, **current** speed, and **turbulence**. In the fast-flowing upper parts

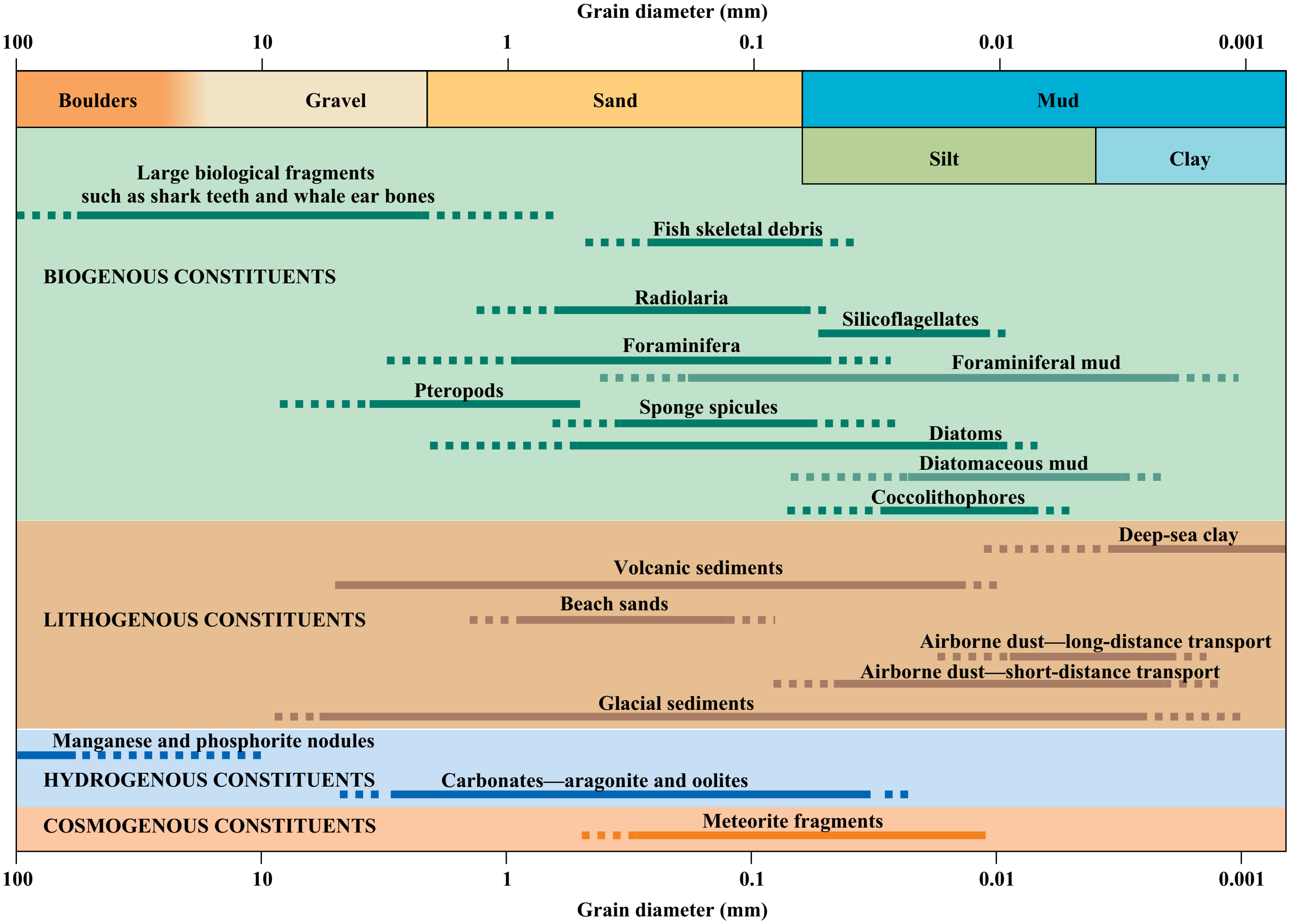


FIGURE 6-1 Various types of particles, each of which has its own characteristic size range, are commonly found in marine sediments. The particles in foraminiferal and diatomaceous muds are, on average, smaller than the sizes of the foraminifera and diatoms from which they are derived because the hard parts of these organisms are partially dissolved and broken mechanically during transport to, and burial in, the sediments

of rivers where they cut through steeply sloped mountain valleys, strong turbulence can be created, and as a result, substantial quantities of rock can be eroded. Such turbulent rivers carry a wide size range of particles. As rivers reach flatter land near the coast, turbulence decreases, and the largest particles are deposited in this region (CC4). Consequently, under normal flow conditions, rivers transport mostly fine-grained particles to the oceans. In contrast, when heavy rains flood rivers, both the volume of the river discharge and the river speed increase dramatically, and great quantities of deposited sediment, including larger particles, are resuspended from the riverbed. River inputs are continuous, but the rate of input, particularly of the larger particles, peaks during major flood events. The peaks can be dramatic. Rivers may transport more sediment to the ocean in a few days after an unusually massive storm, such as a hurricane, than they do during several years or longer of normal conditions.

Not all rivers discharge significant amounts of sediment to the oceans. On many coasts, rivers flow across areas that are now flat coastal plains, although they follow valleys that were cut by glaciers or rivers when sea level was lower. As sea level rose, many of the valleys were inundated to form long **estuaries** in which current speeds are relatively low. Most particles transported by the rivers collect in the estuaries and do not reach the ocean. For example, many of the rivers of the Atlantic coast of North America pass through such estuaries and do not transport much sediment to the ocean. Unless sea level changes, the estuaries will eventually fill with sediment, and deltas may be formed by accumulation of sediment discharged at the river mouths.

Rivers that discharge onto active continental margins at **subduction zones** are generally short and drain limited land areas between the **coastline** and the ridges of the coastal mountain ranges created by the subduction process. The Pacific coasts of North and South America are examples (Fig. 6-2). Because they drain limited land areas, such rivers generally carry only a small load of eroded rock particles. The load may be greater where there are gaps in the coastal mountains, but even so, only small quantities of suspended sediment may be transported to the oceans. For example, the considerable suspended sediment load from the rivers draining the Sierra Nevada in California is deposited in the northern part of San Francisco Bay.

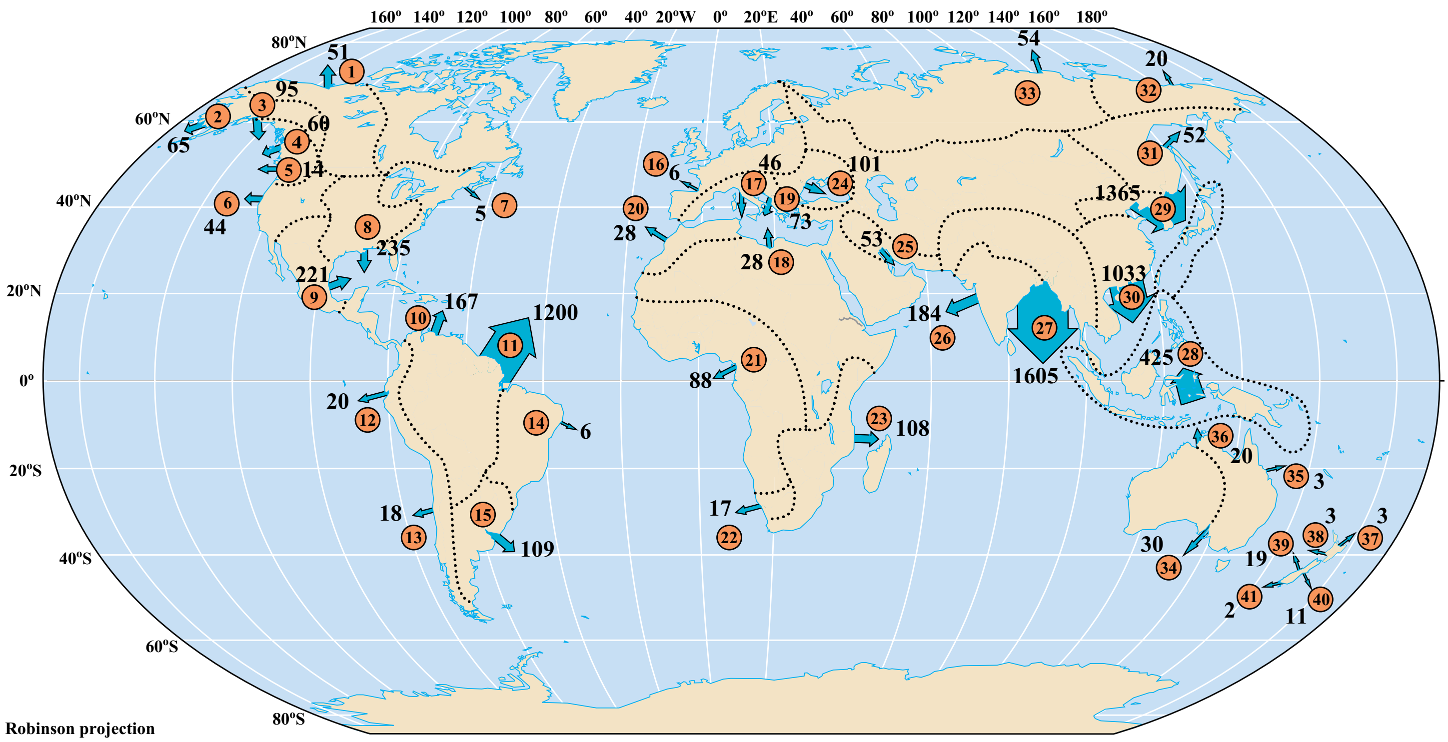
Approximately 90% of all lithogenous sediments reach the ocean through rivers, and 80% of this input is derived from Asia. The largest amounts are from four rivers that flow into the Indian Ocean (Fig. 6-2). The Ganges, Brahmaputra, and Irrawaddy discharge into the Bay of Bengal, and the Indus discharges into the Arabian Sea. Most of the other rivers that transport large amounts of suspended sediment into the ocean empty into **marginal seas**. Examples are the Chang (Yangtze) River (to the East China Sea), the Huang (Yellow) River (to the Yellow Sea), and Mekong (to the South China Sea)—all in Asia—and the Mississippi (to the Gulf of Mexico) in the United States. With the exception of the Amazon's input to the central Atlantic, river discharges of lithogenous sediment to the Pacific and Atlantic Oceans are very limited.

Erosion by Glaciers

Ice is squeezed into cracks in the rocks over which glaciers flow, causing pieces of rock to break off of the floor and sides of the glacial valley. This process of erosion is extremely effective, as can be seen from the deep fjords that the abundant glaciers cut during the last ice age (**Fig. 13.4**). The rock eroded by a glacier is bulldozed, dragged, and carried down the glacial valley with the glacier, and deposited at its lower end. At present, glaciers reach the sea only at certain places in high latitudes, including Antarctica, Alaska, Greenland, and Patagonia. These glaciers release most of the eroded rock into the water close to where they enter the sea. Some icebergs that break off the ends of glaciers may be transported by ocean currents hundreds of kilometers or more before they melt. Rock particles are transported with the ice and can be released far from the glacier, but the amount of sediment transported in this way is small.

Glaciers can transport rock fragments that range from the finest grains to house-sized boulders. The larger boulders, pebbles, and sand grains are almost all deposited in ocean sediments close to the end of the glacier, but finer-grained material can be transported by currents to more remote deposition sites.

Some glaciers release large amounts of "glacial flour," particles so finely ground that they remain suspended for weeks or months, even in lakes or slowly flowing streams where turbulence is very low. Where glaciers discharge into lakes or fjords with long **residence times** (**CC8**), glacial flour can become so



.... Borders of major drainage basins

425 Total sediment transport to ocean (millions of tons per year)

Major rivers in each drainage:

1. Mackenzie, Babbage, Colville
2. Yukon, Kuskokwim
3. Susitna, Copper
4. Fraser, Stekie, Skeena
5. Columbia
6. Russian, Eel, Santa Clara
7. Hudson, Saint Lawrence
8. Mississippi, Brazos
9. Rio Grande, Magdalena
10. Orinoco
11. Amazon
12. Chira, Numerous small
13. Numerous small
14. São Francisco
15. Plata, Negro, Colorado, Parana, Uruguay
16. Rhine, Seine, Loire, Elbe, Garonne
17. Ebro, Tiber, Rhone, Po, Lamone
18. Nile, Medjerda
19. Semani, Drini
20. Several small
21. Niger, Zaire
22. Orange
23. Zambezi, Limpopo, Rufiji, Tana
24. Danube, Dnieper, Don, Coruh, Sakarya
25. Tigris, Euphrates
26. Indus, Namanda
27. Ganges, Brahmaputra, Irrawaddy
28. Sepik, Fly, Purani, Numerous small
29. Huang He (Yellow), Choshui, Kaoping
30. Mekong, Chang (Yangtze)
31. Amur
32. Kolyma, Indigirka
33. Lena, Ob, Yenisey
34. Murray, Darling
35. Burdekin
36. Ord
37. Motu
38. Wanganui
39. Haast, Hokitika
40. Waimkariri
41. Waiau

FIGURE 6-2 Estimated inputs of lithogenous sediments to the oceans by major rivers. These inputs are mostly to marginal seas and are dominated by runoff from the Asian continent to the marginal seas of Southeast Asia. The Amazon is the only very large source of input to the Atlantic Ocean. Inputs to the Pacific Ocean are mostly trapped in the subduction zone trenches that surround it.

FIGURE 6-3 The Cavell Glacier,Mt. Edith Cavell, Jasper National Park, Alberta, Canada. enters a lake. Note the milkiness of the water due to the large quantities of fine particles (glacial flour) released from the melting glacial ice. When glaciers enter the oceans, much of the fine-grained material coagulates because of the high salinity and is deposited quickly. Note also the rock debris on the surface of the glacier itself. This and the dark colored inclusions you can see in the glacial ice provides a profusion of particles of many different sizes. The larger particles will be deposited very close to where the glacier ends.

concentrated that the water is distinctly milky (Fig. 6-3).

Erosion by Waves

Waves continuously **erode** many coastlines (Chap. 11). The eroded rock particles are similar to those introduced to the oceans by rivers. However, particles eroded by waves generally have a larger proportion of unweathered mineral grains, unless the eroding coastline is composed of sedimentary rock (Chap. 11).

Wave erosion of coastlines creates particles of all sizes, from large boulders that fall from undercut cliff faces to the finest clay particles. Waves sort the particles, transporting small ones offshore while leaving larger ones on or close to the shore (Chap. 11).

Transport by Winds

Dust particles can be transported very long distances through the atmosphere by winds before they are deposited on the ground or ocean surface. Dust particles in the atmosphere fall to the ground and are resuspended by air currents in much the same way that particles are deposited and resuspended by ocean currents (CC4). Normal winds can transport only very fine particles, but, as we can observe on a windy day at a dry sand **beach**, high winds can also transport larger grains.

When dust particles fall on the ocean surface, they become suspended sediment in the water. Consequently, there is a continuous flux of dust particles from the land to the oceans. The flux of airborne dust to the oceans is particularly strong in some locations, such as the northern subtropical Atlantic Ocean, where the prevailing winds blow from the Sahara Desert out across the ocean. Dust clouds that stretch thousands of kilometers over the ocean can be seen in this area in satellite images (Fig. 6-4). Filters used by scientists to collect airborne particles in Florida and the Bahamas are often rust-red because they have collected large quantities of the red, clay-sized particles from the Sahara Desert that are transported across the Atlantic Ocean by winds. Windblown dust also is transported from the Gobi Desert in Asia over the North Pacific Ocean. These lithogenous clay-sized particles form a significant fraction of the deep-ocean sediment between 20°N and 30°N in the Pacific Ocean, where the accumulation rate of other particles is very low.

Although the amount of dust in the atmosphere is usually small, the continuous flux of small quantities through the atmosphere adds up to a substantial input to the oceans over geological time. Consider how much dust collects on a piece of furniture in only a few days. Multiply that amount by billions to see how much dust can reach the ocean surface over a period of millions or tens of millions of years.

Fine particles are deposited on all parts of the ocean surface at a relatively uniform rate, although rates of deposition are greater in areas downwind from deserts. High storm winds carry larger particles, up to fine sand size, over the oceans. Such particles are deposited relatively close to their coast of origin because most storm winds do not persist for long periods or blow over great distances.

Although winds generally transport only smaller particles, larger lithogenous particles do enter the atmosphere as a result of volcanic eruptions, especially the explosive eruptions of **convergent plate boundary** volcanoes (Chap. 4). Explosive eruptions can instantaneously fragment and blast upward large quantities of rock, which are then carried by winds as ash and cinders so that they rain out over a very large area. The principal fallout occurs immediately around and downwind of the volcano, where the larger particles drop. Eruptions produce particles of all sizes, including tiny particles that can be carried thousands or tens of thousands of kilometers in the atmosphere before being deposited. The most violent eruptions throw large quantities of ash into the upper atmosphere, where swift winds, such as the **jet stream**, can transport them one or more times around the planet before they finally fall. Volcanic ash from the largest eruptions can remain in the upper atmosphere for years and affect the Earth's climate by reducing the amount of solar energy that passes through the atmosphere to the Earth's surface (CC9). Such ash can also reduce the ozone concentration in the Earth's ozone layer (Chap. 7).

Geologically recent explosive eruptions that have ejected large quantities of ash into the atmosphere include the Mount Pinatubo eruption in the Philippines in 1991 and the Mount St. Helens eruption in Washington in 1980. The Pinatubo eruption completely buried the then U.S. Clark Air Base with ash, even though the base was tens of kilometers from the volcano. The 1980 eruption of Mount St. Helens is estimated to have ejected about 1 cubic kilometer (km^3) of ash into the atmosphere, and Mount Pinatubo about four or five times that amount. These eruptions were dwarfed by two Indonesian eruptions: Krakatau in 1883 and Tambora in 1815, which ejected an estimated 16 and 80 km^3 of ash, respectively. The eruption that created the Long Valley caldera in California 700,000 years ago may have ejected 500 km^3 of ash. Much of the ash ejected by these eruptions would have fallen on the ocean's surface or been washed into the ocean in runoff and then transported to the sediments. Distinct ash layers are found in some ocean sediments corresponding to these and other eruptions.

Transport by Landslides

Landslides occur when loose soil or rock moves down a slope under the force of gravity in a process known as "mass wasting." Water, ice, and winds may be involved in loosening the soil or rock, but the rock or soil is transported by gravity, not by the water, ice, or wind. Some landslides occur on the slopes of a shoreline and may carry rock and soil particles of a wide range of sizes directly into the ocean. Most such material is initially deposited, but it then becomes subject to erosion and transport by waves. Mass-wasting processes also occur on steep slopes on the ocean floor. Slumps or turbidity currents (discussed in more detail later in this chapter) can occur as a result of accumulation of sediment on a continental shelf or at the head of a submarine canyon, or accumulation of volcanic rock on the side of a submerged or partly submerged volcano. These events can transport substantial amounts of lithogenous sediments into the deep-ocean basins.

BIOGENOUS SEDIMENTS

Most non microbial life in the oceans is sustained by the conversion of dissolved carbon dioxide to living organic matter by **photosynthetic** organisms. **Phototrophy** takes place only in the upper part of the water column (at most a few hundred meters deep) where sufficient light is present (Chap. 12, CC14). With only minor exceptions, the photosynthetic organisms of the oceans are small (less than about 2 mm), as are most of the animals that feed on them (Chap. 12). Very few of these organisms live out a full natural life span, because most are consumed by larger animals. Those that are consumed are only partially digested by the larger animals, and their hard parts are not normally decomposed or digested as they pass through the food web. Undigested food particles are packaged together in animal guts to form fecal material that is excreted in the form of fecal pellets. Fecal pellets, although small, are much larger than the microscopic organisms of which they are made up. Therefore, they tend to sink toward the seafloor relatively rapidly, as do bodies of larger animals that have not been ingested by another animal (CC4). Most life in the deep-ocean waters and on the seafloor is sustained by the rain of this organic-rich **detritus** (Chap. 13).

One species, a small **planktonic tunicate** called a "giant larvacean" may contribute a substantial proportion of the total detritus that rains into the deep. The giant larvacean secretes an intricate netlike structure to capture its food. The larvacean lives inside the structure, which, consequently, has been called a "house." When the web becomes clogged with particles, which

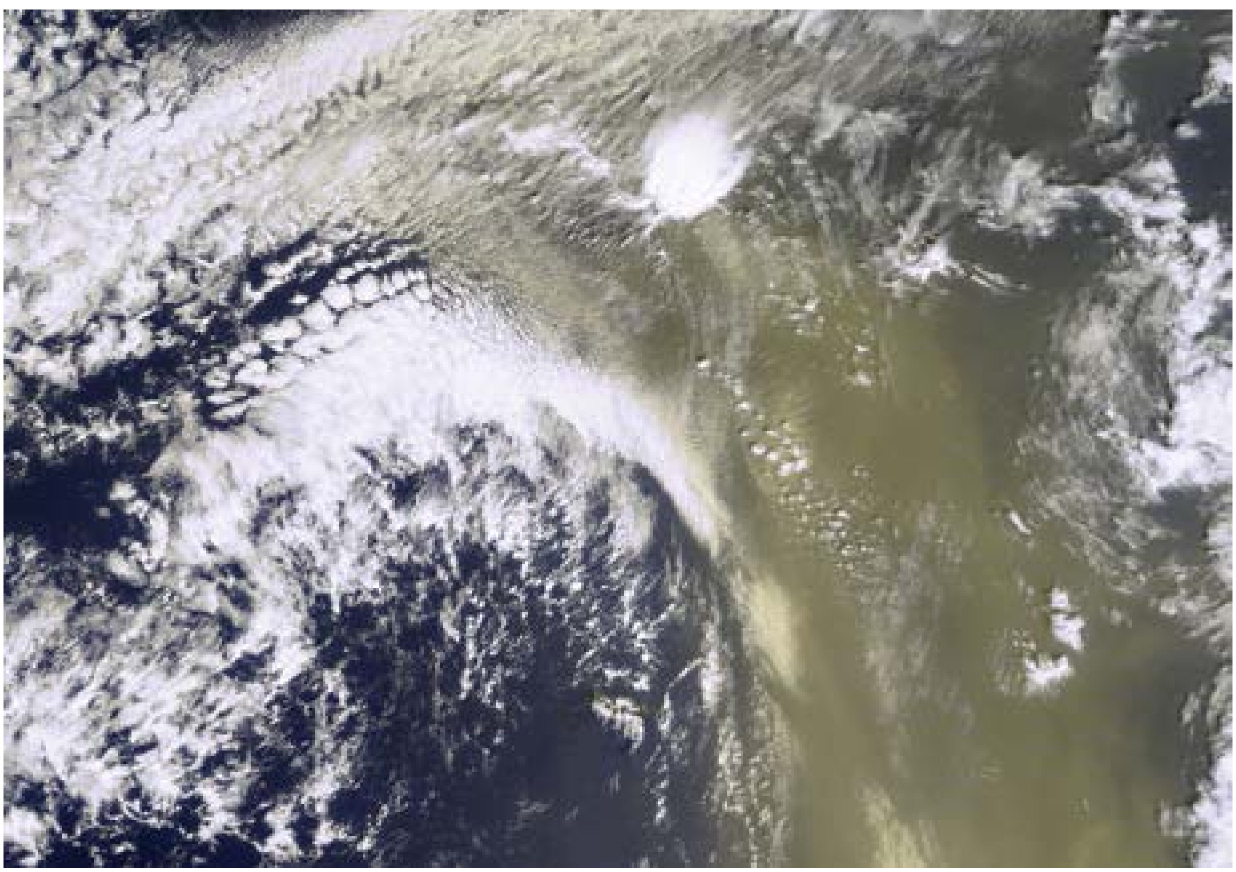


FIGURE 6-4 This August 2004 satellite image shows a huge yellowish-brown, windblown dust plume from the Sahara Desert. The plume extends northward from the African continent. After traveling approximately 1000 km over the Atlantic Ocean, passing over the Canary Islands en route, the plume turns westward toward North America as seen near the top of this image.

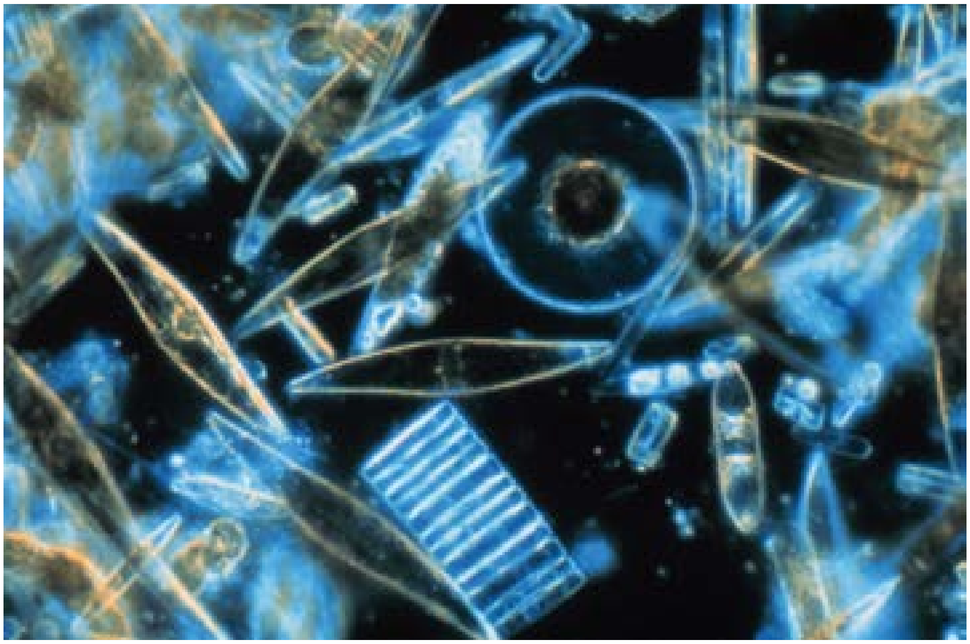


FIGURE 6-5 Several species of diatoms. These are species that were found living between crystals of annual sea ice in McMurdo Sound, Antarctica. Diatoms are algae that are a preferred food source for many small marine animals and animal larvae. They are photosynthetic, and each one is covered by a hard silica frustules

happens about once a day, the larvacean simply releases the house and secretes a new one. The abandoned houses are large enough that they sink fairly rapidly, and this source may then contribute a substantial proportion of the detritus that reaches the seafloor, at least in some areas.

As detritus falls through the water column, it is either utilized by animals or decomposed by **bacteria**, **archaea** and **fungi**, and decomposition continues on the seafloor. In all but a few locations where the rain of detritus is extraordinarily heavy or where there is insufficient oxygen to support bacterial decomposition, the organic matter in detritus is essentially completely decomposed. Consequently, little organic matter is incorporated in the accumulating bottom sediment in most areas. By contrast, many marine species have hard parts that are not decomposed as they fall through the water column, and these materials constitute the overwhelming majority of inputs of biogenous particles to ocean sediment.

The **hard parts**, shells, or skeletons of marine organisms are either **calcareous** or **siliceous.** Calcareous organisms have hard parts of calcium carbonate ($CaCO_3$); siliceous organisms have hard parts of silica in the form of opal, which has the same basic composition as glass (SiO_2). Ocean surface waters are populated by very large numbers of calcareous and siliceous organisms, most of which are smaller than a few millimeters. Hence, a continuous rain of both calcareous and siliceous particles falls from the surface layers. This material does not rain uniformly on the sediment surface, because productivity and types of organisms vary between ocean areas, and because calcium carbonate and silica dissolve in seawater at slow but variable rates.

The two major factors that determine the accumulation rate of biogenous sediment are the rate of production of biological particles in the overlying water column and the rate of decomposition or dissolution of these particles as they fall to the seafloor. The percentage of biogenous material in the sediment is deter-

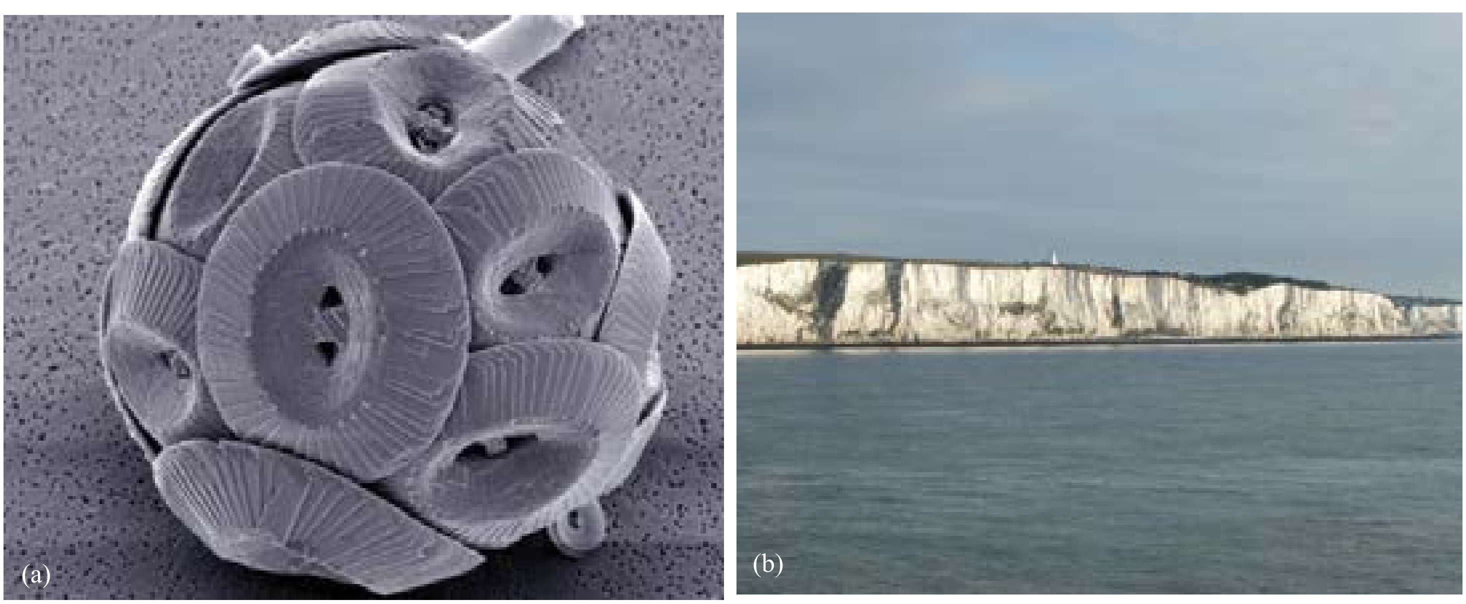


FIGURE 6-6 (a) Coccolithophore (*Coccolithus pelagicus*) photographed at approximately 5000 times magnification. Note the covering of intricate calcareous plates. (b) The rocks that make up the White Cliffs of Dover seen here are composed primarily of coccolithophore plates.

FIGURE 6-7 Zooplankton with calcareous hard parts (a) Most foraminifera are pelagic (live in the water column). However, the species shown in this photograph (the round platelike objects that are about 1 cm in diameter, *Marginopora vertebralis*, Papua New Guinea) is benthic and lives on the coral reef surface. Several of the individuals shown here have died and lost their green pigmentation, revealing the calcareous hard parts that can survive to become sediment particles. (b) A pteropod, or sea butterfly— *Limacina helicina*. The animal "flaps" its "wings" (modified foot) to propel itself through the water. It lives free-swimming in the oceans and eats plankton.

mined by these two factors plus the rate of accumulation of other particles.

Regional Variations of Biogenous Particle Production

Chapters 12 and 13 examine the factors that determine primary productivity rates and the types of organisms that inhabit various regions of the oceans. The distributions of productivity, biomass, and dominant organism type (e.g., calcareous or siliceous) in the ecosystem are major factors in determining the nature of sediments that accumulate in any given area.

In high latitudes and areas of coastal upwelling (Chap. 13), siliceous **diatoms** are the dominant photosynthetic organisms (Fig. 6-5). Diatoms are among the largest of the phytoplankton (up to about 2 mm) and yield relatively large siliceous particles. At lower latitudes and in the open ocean, many of the dominant photosynthetic organisms have no hard parts. However, one group of very tiny photosynthetic organisms called **coccolithophores** (Fig. 6-6a) may grow in abundance. They are covered by a number of small calcareous plates that contribute extremely fine particles to the sediment. The white chalk cliffs of Dover in England consist primarily of coccolithophore plates (Fig. 6-6b).

Zooplankton with calcareous shells include **foraminifera** and **pteropods** (Fig. 6-7), many species of which are present throughout the oceans. Animals with siliceous hard parts are less common, but the microscopic **radiolaria** (Fig. 6-8) have intricate silica shells and are abundant in tropical waters that have high primary productivity.

Dissolution of Biogenous Particles

Seawater is undersaturated with silica. Therefore, siliceous particles dissolve as they fall through the water column to the seafloor. The rate of dissolution is very slow at all depths, but it decreases with depth throughout the upper 2 km of the water column (Fig. 6-9). Thin siliceous hard parts are almost always totally dissolved and reach the sediment only in areas where siliceous organisms grow in great abundance. For example, many radiolaria have thin, intricate shells that are dissolved relatively easily and reach the sediment in substantial quantities only in highly productive tropical waters where they abound.

The production of **fecal pellets** can enhance the accumulation of siliceous and calcareous particles. When packaged in relatively large fecal pellets, such particles fall more quickly to the seafloor and are partially protected from dissolution until the organic matter in the pellet disintegrates.

Many diatoms have a relatively thick siliceous frustule (hard part), much of which reaches the seafloor, where it continues to dissolve until buried by other particles. Diatom frustules are present in the sediment in locations where their production rate and the rate of sediment accumulation (the sedimentation rate) are high.

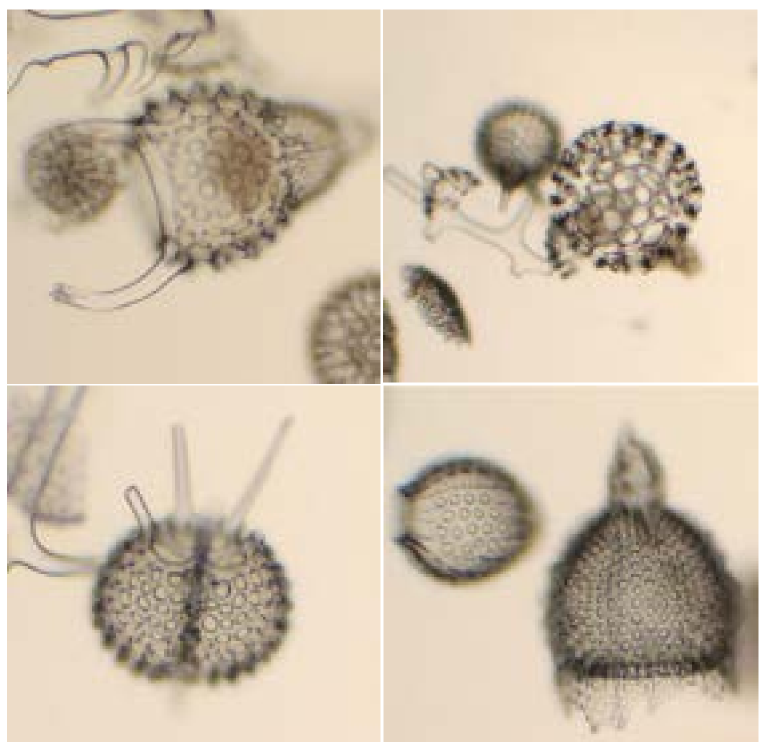


FIGURE 6-8 Radiolaria (photographed here at about 10 times magnification) are small organisms that have intricate silica shells, which make a major contribution to the sediments. They occur in abundance in some tropical waters where primary productivity is high in the surface water layer.

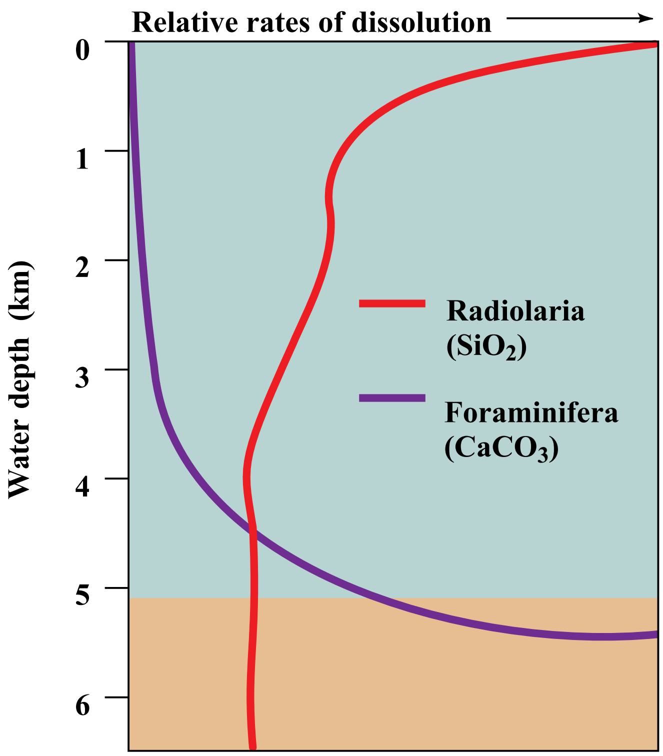


FIGURE 6-9 Changes in the dissolution rate of siliceous shells (radiolaria) and calcareous shells (foraminifera) with depth. Siliceous shells are more soluble in surface layers; calcareous shells are more soluble at greater depth.

Calcareous particles also dissolve in seawater, but their behavior is different and more complex than that of siliceous particles. In the upper water layers, where water temperatures are high and pressures low (**Chap. 8**), seawater is generally supersaturated with calcium carbon carbonate. Consequently, calcareous shells dissolve only very slowly or not at all in near-surface waters (**Fig. 6-9**). The solubility of calcium carbonate increases with pressure and with decreasing temperature. Therefore, solubility tends to increase with depth.

Not all calcareous hard parts are made of the same mineral. Calcium carbonate shell material is of two distinct forms: **calcite** and **aragonite**. Some types of animals, including most foraminifera, have calcite shells, whereas others, such as pteropods, have aragonite shells. Aragonite dissolves much more readily in seawater than calcite. Therefore, pteropod shells are dissolved completely at shallower depths than foraminiferal shells. Where pteropods are more abundant than foraminifera, shallow-water sediments can consist of predominantly pteropod particles. Sediments at intermediate depths are predominantly foraminiferal, and sediments at greater depths have almost no calcareous component (**Fig. 6-10**).

Carbonate Compensation Depth

Below a certain depth, seawater is undersaturated with respect to calcium carbonate, and calcareous debris starts to dissolve. Below this level, the degree of undersaturation and hence the dissolution rate of calcium carbonate increase with depth. Eventually, at the **carbonate compensation depth** (**CCD**), the dissolution rate is fast enough to dissolve all of the calcium carbonate before it can be incorporated in the sediment.

The CCD depends not only on pressure and temperature, but also on other factors, especially the concentration of dissolved carbon dioxide. An increase in carbon dioxide concentration lowers the pH, making the water more acidic, and thus increases calcium carbonate solubility.

Carbon dioxide solubility in seawater increases as temperature decreases. The deepest water layers of the oceans were formed by the sinking of cold water in certain high-latitude regions (**Chap. 8**). Thus, the deepwater layers contain relatively high concentrations of dissolved carbon dioxide. In addition, deep waters below the warm surface water layer accumulate dissolved carbon dioxide as animals respire and bacteria and other decomposers convert organic matter to energy and carbon dioxide

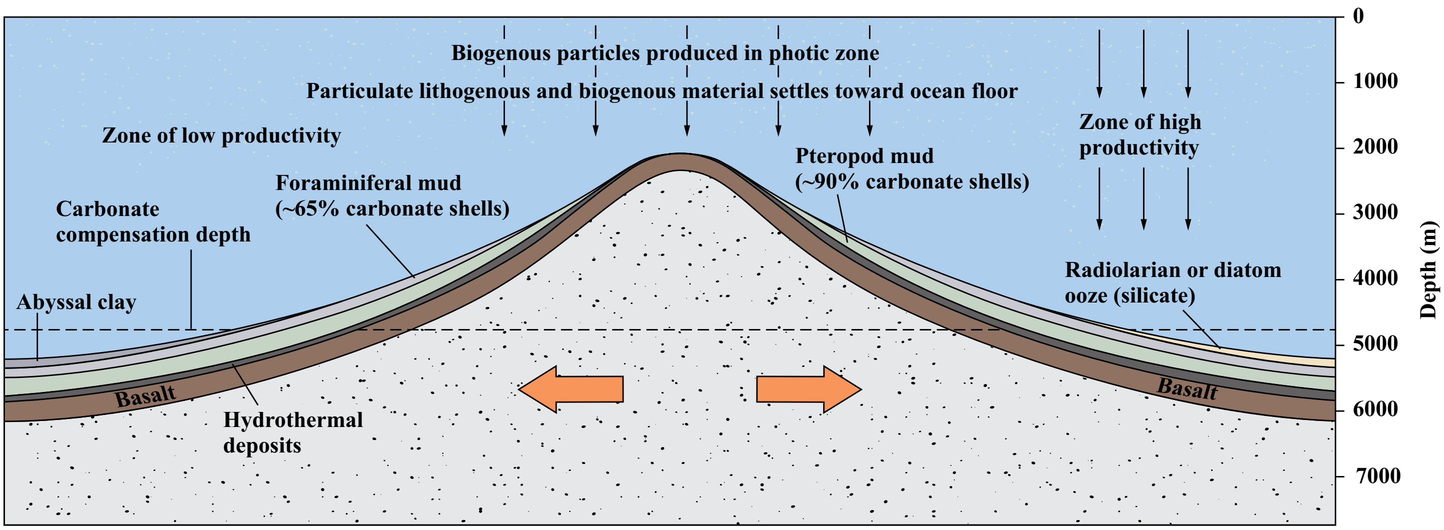


FIGURE 6-10 Idealized depiction of layered sediment accumulation on an actively spreading oceanic ridge. There is a thin, patchy cover of predominantly hydrothermal sediments at the center of the oceanic ridge. On the flanks of the ridge, which are shallower than the carbonate compensation depth (CCD), calcareous sediments dominated by pteropods and foraminifera are deposited. However, pteropods dominate the surface sediments only at depths much shallower than the CCD because their shells are made of a more soluble form of calcium carbonate than those of foraminifera. As the new oceanic crust moves away from the spreading center and sinks isostatically, it is covered by layers of hydrothermal sediments, pteropod and foraminiferal oozes, and then siliceous sediments. These siliceous sediments can be either biogenous, if diatomaceous or radiolarian production in the overlying water column is high, or fine-grained lithogenous clays. In the Pacific Ocean, the lithogenous sediments are primarily deep-sea clays because subduction zones prevent inputs of turbidites. In the Atlantic Ocean, they are generally a mixture of deep-sea clays and turbidites.

(**Chap. 12**). Therefore, older deep water (water that has been away from the surface longer) has a higher carbon dioxide concentration than newer deep water. The deep water of the Pacific Ocean is older than the deep water of the Atlantic Ocean (**Chap. 8**). Consequently, calcium carbonate is more soluble and the CCD is shallower in the Pacific Ocean because of its higher dissolved carbon dioxide concentrations. The CCD is at approximately 4000 m in the Atlantic, at about 2500 m in the South Pacific, and at less than 1000 m in the North Pacific, which has the oldest deep waters with the highest carbon dioxide concentrations.

Anthropogenic Carbon Dioxide and the Carbonate Compensation Depth

The carbonate compensation depth (CCD) does not remain fixed but changes with climate, particularly with changes in the carbon dioxide concentrations of ocean water. An understanding of these changes is necessary to predict the fate of the excess carbon dioxide released to the atmosphere by the burning of fossil fuels (**Chap. 7**, **CC9**).

Atmospheric carbon dioxide, carbon dioxide dissolved in seawater (primarily as carbonate and bicarbonate ions), and carbon locked up in calcareous sediments or sedimentary rocks are all involved in the same biogeochemical cycle. Some of the excess carbon dioxide released to the atmosphere has been absorbed by ocean water and some may have been captured in calcareous sediments. However, the higher carbon dioxide concentrations lower seawater **pH** and will move the CCD to shallower depth, causing more calcium carbonate to dissolve (**Chap. 5**). The effects of this ocean **acidification** on marine organisms in the surface layers of the oceans are a primary concern. However, as more calcium carbonate dissolves, this will result in even higher concentration of carbonate and, therefore, carbon dioxide in the deep layers of the oceans. Eventually, this will cause more carbon dioxide to be released to the atmosphere when the deep waters are brought to the surface. How fast this will take place is, as yet unknown. However, because atmospheric carbon dioxide and the CCD have varied substantially throughout the Earth's history, we may be able to find an answer to this and other questions about the effects of ocean acidification in the record preserved in deep-sea sediments. Indeed, existing evidence suggests that release of carbon dioxide to the atmosphere from carbon dioxide rich deep ocean waters may have caused or contributed to major warming events in Earth's history such as the interglacial periods of the last **ice age**.

HYDROGENOUS SEDIMENTS

The input of hydrogenous sediments to the ocean floor is small in comparison with the inputs of biogenous and lithogenous material. Nevertheless, hydrogenous material can be an important component of the sediment, particularly in some areas where the accumulation rate of biogenous and lithogenous material is low.

Seawater is generally undersaturated with most dissolved substances. Consequently, hydrogenous sediments are formed under special conditions where the chemistry of ocean water is altered. Various types of hydrogenous sediment form under different sets of conditions. The types include hydrothermal minerals, manganese nodules, phosphorite nodules and crusts, carbonates, and **evaporites**.

Hydrothermal Minerals

Because the Earth's crust is thin at the **oceanic ridges**, more heat flows through the seafloor from the **mantle** at the ridges than elsewhere. The excess heat drives a series of **hydrothermal vents** that discharge hot water (sometimes several hundred degrees Celsius) into the cold surrounding seawater. The hot water is devoid of dissolved oxygen when discharged, and it contains high concentrations of metal sulfides, such as iron sulfide and manganese sulfide. These sulfides are soluble in the absence of dissolved oxygen, but they are oxidized to form insoluble hydrous oxides as the vent plumes mix with much larger volumes of the surrounding oxygenated seawater. The hydrous oxides precipitate to form a cloud of fine, metal-rich particles. These particles sink to the seafloor and accumulate as metal-rich sediment in the area surrounding the vent.

In addition to iron and manganese, hydrothermal vent deposits contain high concentrations of other metals, including copper, cobalt, lead, nickel, silver, and zinc. Although some particles accumulate to form metal-rich sediment near the vents, many fine particles formed at the vents are probably transported large distances before settling and provide an input of hydrogenous particles to sediments throughout the deep-ocean basins. This may also be the origin of the material that forms the manganese nodules described later in this chapter.

The mechanism of hydrothermal vent circulation is not fully understood. The excess heat flow at the ridge is believed to drive the mechanism. Water within the rocks and sediments beneath the hydrothermal vents is heated and rises (**CC1**) through the vents into the water above. This rising water is replaced by cold seawater that seeps through cracks in the rocks and sediment at other locations nearby and is transported to the vent through the rocks and sediment (**Chap. 15**). As the seawater migrates to the vent through the sediment and rocks, it is heated, its oxygen is depleted by reaction with sulfides, and its pH drops so that it leaches salts and metal sulfides from the sediment and surrounding rock before it is discharged at the vent. The vents are, in essence, parts of convection cells (**CC3**, **Fig. 15.16**).

Hydrothermal vents were not discovered until the late 1970s. They are surrounded by commercially attractive deposits of metal-rich sediment, and they also support unique communities of organisms, many of which are found nowhere else on the Earth (**Chaps. 12, 15**). Hydrothermal vents are limited to small areas on the ridge, often separated by substantial distances, but they are present throughout the oceanic ridge system. Whether the metal-rich sediment is abundant enough to be commercially exploitable is not known. In addition, whether the unique biological communities could be safeguarded if the minerals were commercially exploited is uncertain.

Along the axis of the Red Sea, where a new oceanic ridge is forming, hydrothermal vents have caused huge quantities of hydrothermal minerals to collect. The water in the deep basins of the Red Sea is isolated from and does not mix with open-ocean waters. As a result, hot water discharged by the hydrothermal vents has accumulated in the Red Sea deep basins because, even though the water is hot, it is also very high salinity and thus high density (**CC6**). The high density strongly inhibits vertical mixing of this deep water with Red Sea surface water so the hot, high salinity, deep water is also **anoxic** which prevents the minerals discharged by the hydrothermal vents from being oxidized. These factors have caused metal rich **hydrothermal minerals** to accumulate in large quantities in sediments and high concentrations of suspended sediments in these Red Sea deep water areas. The deposits are so extensive that test mining was performed in 1979

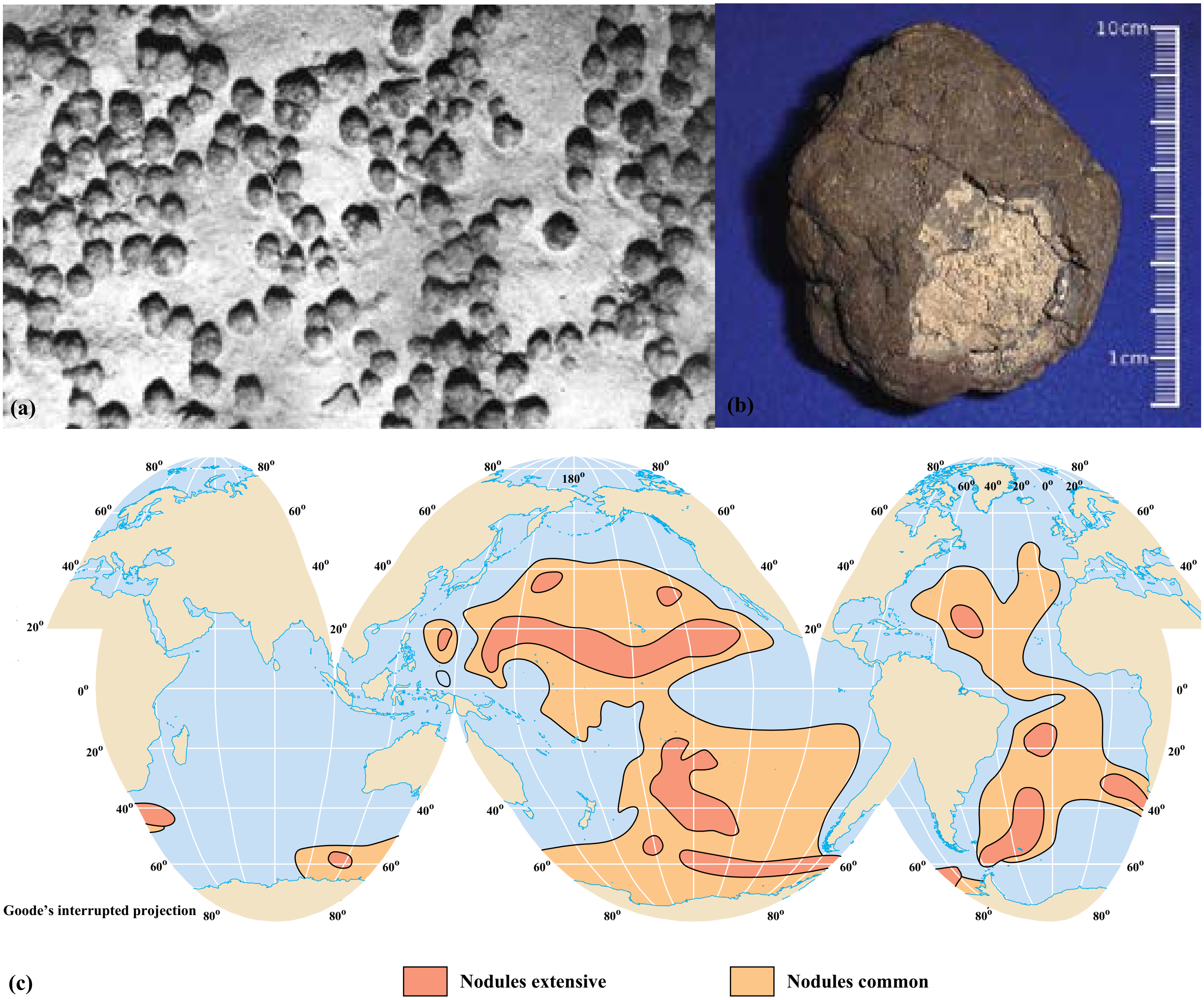


FIGURE 6-11 (a) Manganese nodules cover certain areas of the deep-ocean floor, such as this area of the northeastern Atlantic Ocean. (b) A typical manganese nodule. The nodule is made up of concentric layers. (c) The areas of the greatest density of manganese nodules are on the abyssal floor of the North Pacific Ocean between about 10°N and 20°N and in the central South Pacific Ocean.

but no further mining has occurred yet as the minerals are still available at lower cost from mining on land. Nevertheless, these hydrothermal minerals that have collected on the Red Sea floor may eventually represent an economic resource as mineral scarcities grow in the future.

Undersea Volcano Emissions

Undersea volcanic activity is not limited to the oceanic ridges. Volcanoes also erupt under the sea at some hot spots and at some convergent plate boundaries, such as the Mariana Arc. Sulfide-rich water, comparable to the hydrothermal fluids discharged at oceanic ridge vents, is discharged into the water column at some locations on these undersea volcanoes. The sediments in the areas surrounding these discharges sustain populations of biological communities similar to those found surrounding oceanic ridge hydrothermal vents, and these sediments are likely to be similarly rich in some metals. There is particular interest in studying these communities and discharges because these undersea volcanoes are located at much shallower depths than most oceanic ridge hydrothermal vents. This makes them easier to reach for study. Even more importantly, some of these shallower vents are different because they discharge into the upper layers of the ocean where photosynthesis also occurs. This raises many questions about how these relatively shallow discharges and communities may affect and/or interact with the chemistry and biology of the surrounding waters.

Manganese Nodules

Manganese nodules are dark brown, rounded lumps of rock, many of which are larger than a large potato. Enormous numbers of nodules litter parts of the deep-ocean floor (Fig. 6-11a). They are potentially valuable as a mineral resource because they usually consist of about 30% manganese dioxide, about 20% iron oxide, and up to 1% or 2% other metals that are much more valuable, including copper, nickel, and cobalt.

Manganese nodules are probably formed by various mechanisms and from various sources of manganese, iron, and other elements. These chemicals may come from the dissolved ions in river water, from seawater leaching of volcanic materials in the oceans, or from the hydrothermal vents on the oceanic ridges.

At present, the major source is thought to be the oceanic ridge hydrothermal vents. Particles of colloidal size (submicroscopic) are hypothesized to form at the hydrothermal vents, be transported throughout the oceans, and accumulate by adsorption on the surface of the nodule, perhaps aided by microbial action.

Most nodules initially form around a large sediment particle (e.g., a shark's tooth) and are then built up in layers (**Fig. 6-11b**) at the very slow rate of about 1 to 10 mm per million years. Nodules are most common where the sedimentation rate is extremely slow. In areas where sediment accumulates rapidly, nodules are buried by new sediment before they have time to grow. Where nodules do form, occasional disturbance of the nodule by marine organisms is hypothesized to be necessary to prevent the nodule from being buried.

The areas of greatest accumulation of manganese nodules are the deepest parts of the Pacific Ocean (**Fig. 6-11c**), where the sedimentation rate is very low. The greatest density of manganese nodules is in the region of the Pacific Ocean south of a line between Hawaii and southern California and north of 10°N latitude.

Phosphorite Nodules and Crusts

Phosphorite is a mineral composed of up to 30% phosphorus. **Phosphorite nodules** form in limited areas of the continental shelf and continental slope and on some seamounts. Phosphorite nodule formation apparently requires low dissolved oxygen concentrations in bottom waters and an abundant supply of phosphorus. These conditions are present in upwelling regions where productivity is high. In such regions, the decomposition of falling detritus depletes oxygen in the bottom waters and supplies relatively large quantities of phosphorus, which is released as phosphate when the organic matter decays.

Phosphorite nodules grow slowly (1 to 10 mm per 1000 years). In contrast to manganese nodules, they do not form concentric layers but instead grow only on the underside, accumulating phosphate released by the decomposition of organic matter. Because very large deposits of phosphates are available on land, phosphorite nodules are unlikely to become commercially valuable.

Carbonates

Many **limestone** rocks lack **fossils.** Some consist of biogenous sediments that have lost their fossils through **diagenetic** changes (discussed later in this chapter), but others consist of hydrogenous sediments formed by direct precipitation of calcium carbonate. Calcium carbonate precipitates from seawater under conditions that apparently were widespread in the oceans at different times in the past, but now are present only in very limited regions, such as the Bahamas.

Calcium carbonate precipitation is more likely when water temperatures are high, because calcium carbonate solubility decreases with increasing temperature, as well as when the concentration of dissolved carbon dioxide is lowered, which raises the pH (**Chap. 5**). Carbon dioxide concentrations can be reduced by high rates of photosynthetic production (**Chap. 12**, **CC14**). They can also be reduced when high temperatures reduce the carbon dioxide solubility of surface waters and allow some dissolved carbon dioxide to escape into the atmosphere.

In the present-day oceans, both high temperature and high productivity appear to be necessary to reduce the calcium carbonate solubility enough to cause precipitation. For example, warm water flowing north through the Straits of Florida (**Chap. 8**) is upwelled across the shallow Bahama Banks (**Fig. 6-12**). Over the banks, it is heated further and sustains high primary productivity because it is nutrient-rich and sunlight is intense (**Chap. 12**, **CC14**). This causes the pH to rise, and as a result, calcium carbonate precipitates around suspended particles to form white, rounded grains called "ooliths." Ooliths collect in shallow-water sediment in many areas. Where they are especially abundant in relation to other particles, the sediments are called "oolite sands."

Evaporites

In marginal seas in arid climates, the evaporation rate of ocean water exceeds the rate of replacement by rainfall and rivers (**Chap. 7**). Under such conditions, salinity progressively increases until the seawater becomes saturated with salts. As evaporation continues, salts, called "evaporites," are precipitated in succession: calcium and magnesium carbonates first, then calcium sulfate, and finally sodium chloride. Only a few marginal seas are evaporating in this way today, and even in these areas salt precipitation occurs only on tidal flats or in shallow embayments around the perimeter of the sea. Areas where evaporites form today include the Dead Sea, Red Sea, and Persian Gulf. Such seas must have been more abundant in the past because salt layers are present at many locations deep within ocean sediments, as well as in previously submerged parts of the continents.

Seismic profiles reveal several extensive layers of buried evaporite sediments, some more than 100 m thick, beneath the floor of the Mediterranean Sea. For such layers to have formed, the entire Mediterranean must have evaporated almost to dryness. This evaporation process apparently occurred several times

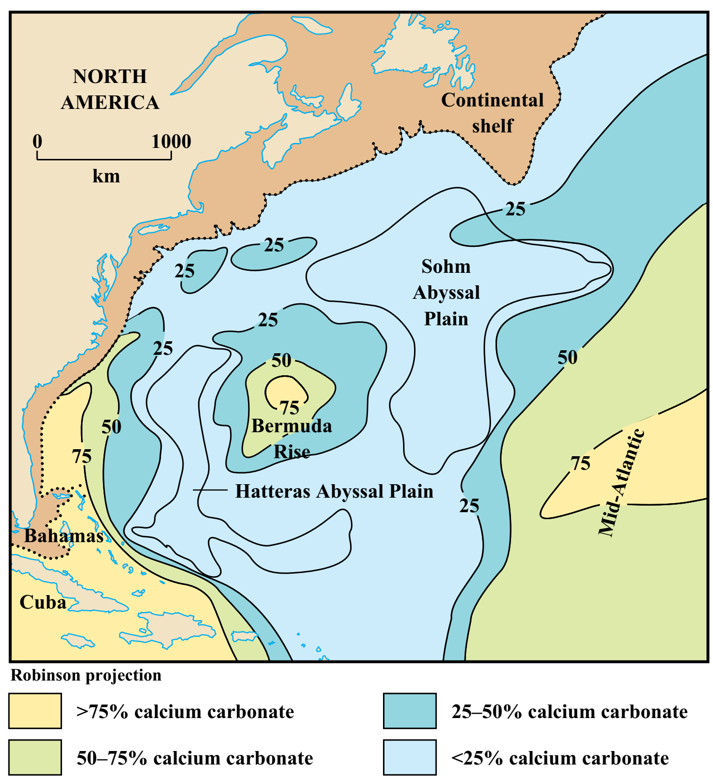


FIGURE 6-12 Calcium carbonate distribution in the sediments of the western North Atlantic. Values are the percentage of calcium carbonate in the sediment. The high percentages of calcium carbonate on the Bahama Banks are due primarily to the dominance of hydrogenous carbonate sediments. The shallow parts of the Bermuda Rise and the Mid-Atlantic Ridge are regions where lithogenous sediment accumulation rates are low, allowing biogenous carbonate particles to dominate the sediments in these regions.

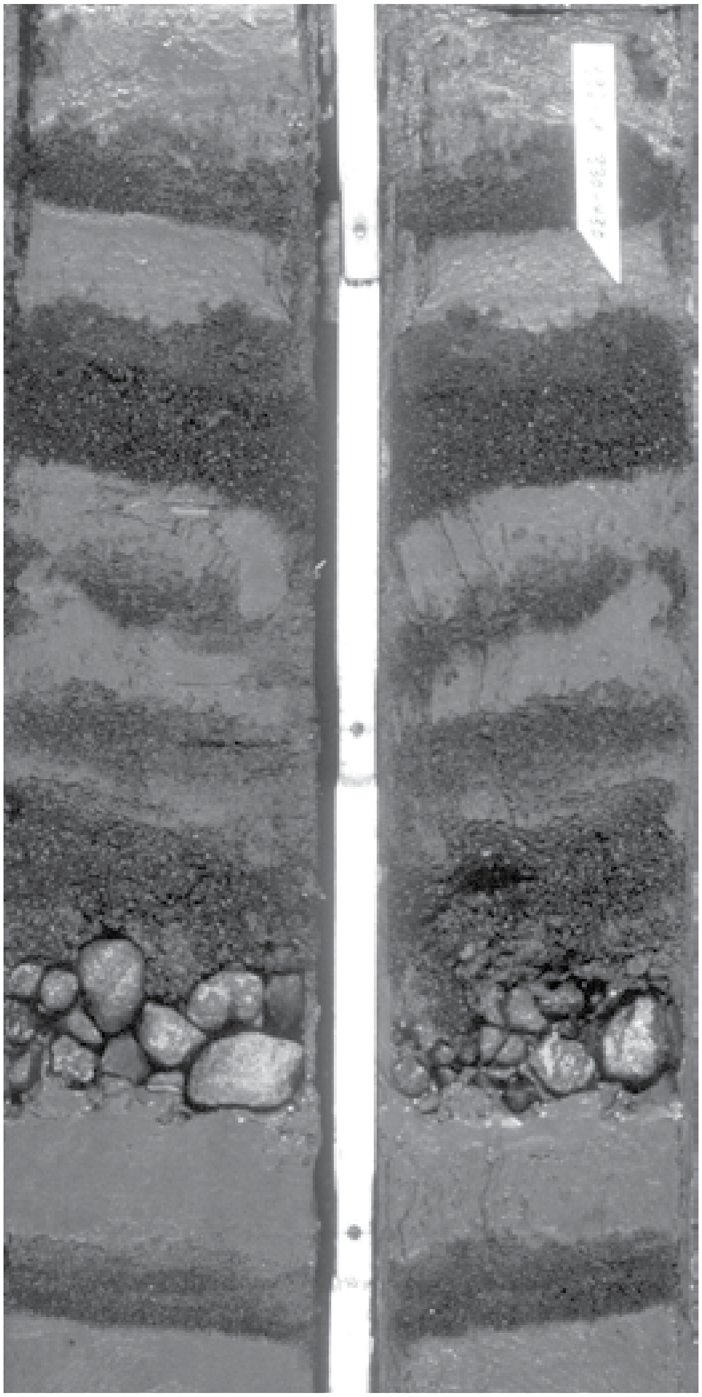


FIGURE 6-13 There are a number of turbidite layers in these cores. Note the sharp change in grain size at the bottom of each turbidite layer and the gradual reduction in grain size above this boundary. The convex shape of the boundaries between layers is due to distortion of the core during the drilling process. Most turbidte layers would not exhibit the very large particles seen at the lower part of some of the turbidite layers. However, these cores are from the Antarctic continental slope where large particles are carried onto the continental shelf by glaciers

when global sea level was lower and the connection between the Mediterranean and the Atlantic Ocean was broken at the sill that now lies underwater across the Straits of Gibraltar.

COSMOGENOUS SEDIMENTS

Most meteorites that enter the Earth's atmosphere are destroyed before reaching the surface. As they burn up, the meteorites form particles that rain out onto land and ocean and are eventually incorporated into sediments. These cosmogenous particles contribute only minor amounts of material to ocean sediments in comparison with biogenous, lithogenous, and hydrogenous sources. Nevertheless, the total input of meteorite dust to ocean sediments is estimated to be tens of thousands of tonnes each year.

Two types of meteorites hit the Earth: stony and iron-rich. The dust from the meteorites is found in sediment as cosmic spherules, which are spherical particles that show the effects of having been melted in the atmosphere. Stony meteorites form silicate spherules that are difficult to distinguish from lithogenous sediment grains. Most iron-rich spherules are about 0.2 to 0.3 mm in diameter and are magnetic, which makes them easier to find.

Recent studies have focused on finding the impact craters and debris from large meteorites that were not destroyed in the atmosphere but instead struck the oceans. A particularly large meteorite may have landed in the Caribbean Sea or Gulf of Mexico 65 million years ago. That impact is discussed later in this chapter.

SEDIMENT TRANSPORT, DEPOSITION, AND ACCUMULATION

Sediment particles are transported by ocean currents and waves in the same way that dust and sand are blown around by winds. High winds and fast water currents both cause particles to be suspended and carried until the wind or current speed diminishes. Particles are then deposited on the ground or ocean floor, but they may be picked up again if the wind or current speed increases. CC4 explores the relationship between particle size and sinking rate, between current speed and sinking rate, and between current speed and resuspension of deposited particles.

Large particles sink rapidly, and high current speed is necessary to prevent them from being deposited. Once deposited, large particles are not resuspended unless current speed is considerable. Many large particles are transported to the ocean by rivers during periods of peak river flow that follow flooding rains. When such particles reach the ocean, where currents are weaker, they are either deposited or resuspended and transported by waves. In areas where wave energy is limited, the large particles are not resuspended, but they remain in sediment at the river mouth.

Waves can generate orbital water motions that have higher speeds (**orbital velocity**) than ocean currents have. As a result, they can resuspend large particles in waters that are shallow enough for the wave energy to reach the seafloor, and sand-sized particles can be resuspended and transported in the nearshore zone. Waves can transport sand long distances along the coast in shallow water (Chap. 11). However, the speed of water motion in waves is reduced with depth below the surface (Chap. 8). Sand-sized particles that are transported offshore are deposited when they reach depths at which the wave orbital velocity is no longer high enough to resuspend them. Smaller sand-sized particles that can be resuspended at lower water speeds are deposited in deeper waters than larger particles. Thus, waves tend to sort sand-sized particles in nearshore sediments: larger particles in shallow water and progressively smaller particles with increasing water depth.

Clay-sized particles are generally **cohesive**, but they tend to remain in suspension once resuspended, unless current speed is reduced substantially (CC4). The peak water velocity in waves is often sufficient to resuspend most cohesive sediments, and fine particles accumulate in nearshore areas only if these areas are

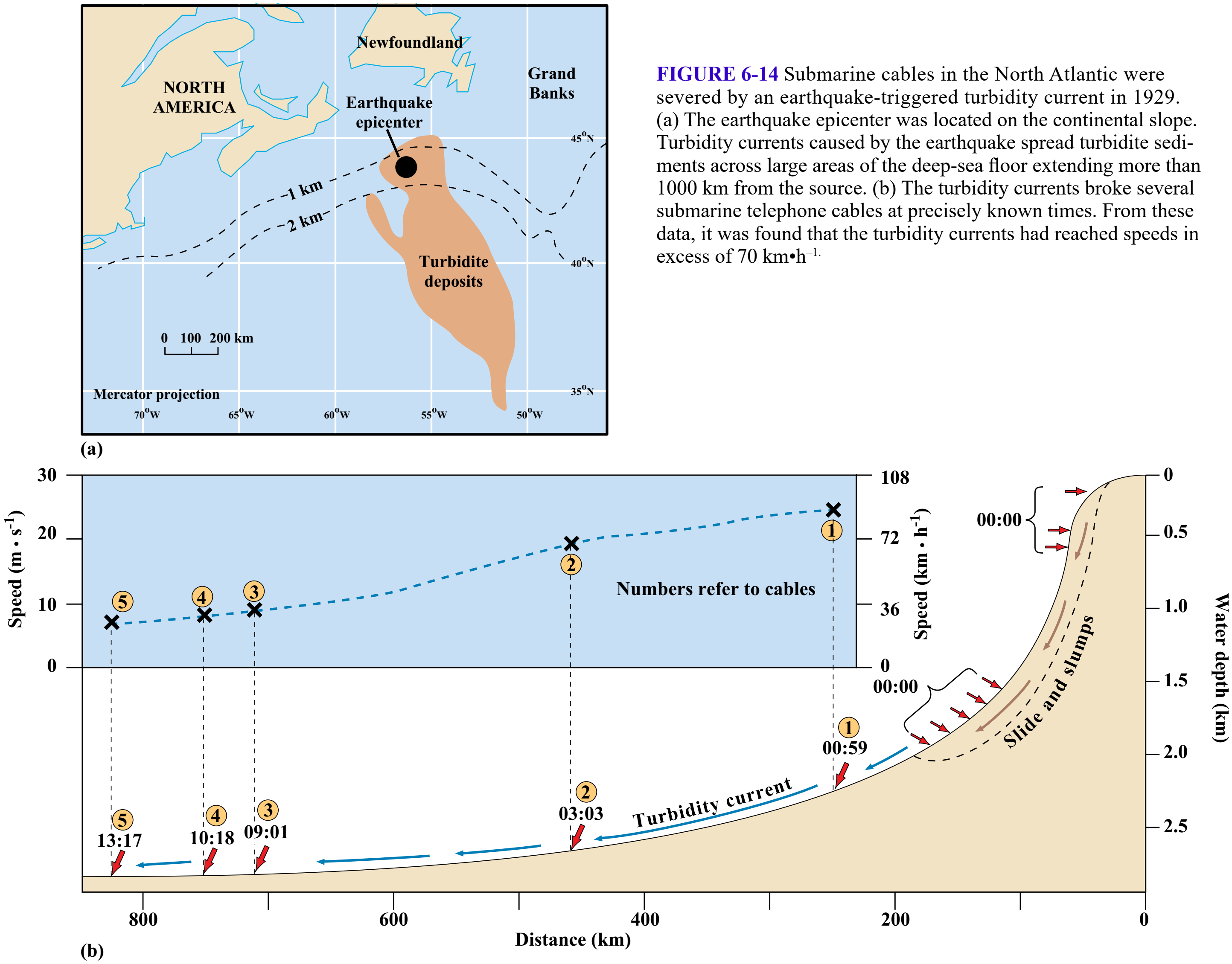


FIGURE 6-14 Submarine cables in the North Atlantic were severed by an earthquake-triggered turbidity current in 1929. (a) The earthquake epicenter was located on the continental slope. Turbidity currents caused by the earthquake spread turbidite sediments across large areas of the deep-sea floor extending more than 1000 km from the source. (b) The turbidity currents broke several submarine telephone cables at precisely known times. From these data, it was found that the turbidity currents had reached speeds in excess of 70 km•h^{-1}.

well protected from waves and have slow currents. Such areas include both wetlands and fjords. Fine particles are transported by currents until they reach a low current area, where they are deposited permanently. The finest particles may be transported many thousands of kilometers and for many years before being deposited in the deep oceans, but they are often combined into clumps by electrostatic attraction or packaged into fecal pellets. The larger conglomerated particles are deposited more rapidly.

Aside from an occasional large particle (e.g., shark's tooth or whalebone), deep-ocean sediments remote from coastal sediment inputs are fine-grained, because almost all large particles are deposited in nearshore sediments and most particles introduced directly to the deep ocean are small. For example, most marine organisms are microscopic and therefore produce small particles. Meteorite dust particles, dust particles carried by winds, and most hydrogenous particles are also small.

Turbidity Currents

In some deep-ocean sediments, particularly those on abyssal plains adjacent to continental slopes, layers of coarse-grained sediments (sand and gravel) can be found. Layers of such sediments are separated by layers of the fine-grained sediments that normally accumulate on the deep-ocean floor. The coarse-grained sediments are transported downslope to the abyssal plain by **turbidity** currents.

Turbidity currents are similar to avalanches. Snow accumulates on a mountainside until it becomes unstable, breaks loose, and tumbles down the slope as an avalanche. Similarly, sediments accumulate on the continental shelf edge and slope until they become unstable, break loose, and flow down the continental slope as turbidity currents. Turbidity currents can be triggered by disturbances such as earthquake vibrations or sudden large discharges of sediment by rivers in the same way that avalanches can be triggered by noises or storm winds. Pockets or layers of methane or methane hydrates (Chap. 16) formed by the decay of organic matter in the sediments may rupture and be released when a turbidity current occurs. This may have the effect of providing a "lubricated" layer on which the turbidity current travels, which may enhance both the size and speed of the turbidity current.

Once a turbidity current has been triggered, the disturbed sediments flow down the continental slope. As it gathers momentum, the turbidity current entrains more sediment and water, just as an avalanche gathers more snow.

Turbidity currents can flow down the continental slope at speeds of 70 km·h^{-1} or more, which are sufficient to suspend and retain large particles in suspension. The speed is reduced when the turbidity current reaches the abyssal plain, and the entrained sediment particles are deposited from the suspended sediment cloud. The larger particles are deposited first, followed by progressively smaller particles, as speed and turbulence diminish. The result is the formation of a graded bed of sediments on the abyssal seafloor within which the largest grains are at the bottom and grain size progressively decreases upward (**Fig. 6-13**). Such graded beds, called "**turbidite** layers," can be meters thick. In many locations, a number of them are separated by layers of finer sediment. The fine-grained sediment layers were deposited slowly in the years, centuries, or millennia between successive turbidity currents.

Turbidity currents can travel long distances on the abyssal plain before finally depositing all their terrigenous sediment load. Turbidity current deposits are generally thickest at the lower end of submarine canyons, where they form abyssal fans that decrease in thickness and median grain size in a seaward direction.

Where a deep-ocean trench is present at the base of the continental slope, turbidity currents are intercepted by the trench and do not reach beyond it to the abyssal plain. This is one reason why the topography of the Atlantic Ocean abyssal floor is flatter than that of the Pacific Ocean. The Atlantic Ocean floor topography is buried by turbidites that have been able to reach the oceanic ridge since the Atlantic Ocean first began to form. Because most of the Pacific Ocean floor was created after subduction zones formed around this ocean, turbidite deposits in the Pacific are rare.

Turbidity currents must be common events in the oceans, particularly where continental shelves are narrow and terrigenous sediment inputs are high. They are difficult to observe because they last only a few hours at most and occur unpredictably, although some cause tsunamis. Nevertheless, their destructive power must be respected. In fact, this destructive power is what led to the first quantitative observation of the speed and geographic extent of turbidity currents. In 1929, an earthquake occurred in the Atlantic Ocean off Nova Scotia. The turbidity current triggered by the earthquake plunged down the continental slope, and snapped and buried several undersea telephone cables (**Fig. 6-14**). Because the precise times at which successive cables broke were known, the peak turbidity current speed was later estimated to be at least 70 $km \cdot h^{-1}$. This turbidity current, like many others, moved primarily down a submarine canyon and traveled more than 600 km before slowing and depositing a turbidite layer across the abyssal plain. Because turbidity currents often flow down submarine canyons, it has been suggested that their scouring action maintains or even creates the canyons.

By studying the layers of sediments, scientists have determined that extremely large turbidity currents have occurred in the past and, therefore, may occur again. For example, about 20,000 years ago, a turbidity current occurred in the western Mediterranean that was estimated to have deposited 500 km^3 of sediments on the deep-sea floor, enough material to cover all of Texas with nearly 2 m of mud and sand. Another slide off the coast of Norway occurred 30,000 to 50,000 years ago, involved more than twice the volume of sediment as the Mediterranean example, and left a scar in the continental slope that is larger than the state of Maryland.

Turbidity currents carry shallow-water organisms to great depths, where some, particularly microorganisms, may survive and adapt. Their occurrence also may affect the food web. After the 1989 Loma Prieta earthquake in California, dense schools of fish were observed feasting on the sediment-dwelling organisms exposed on the floor and sides of Monterey Canyon, where the sediment slumped and presumably caused a turbidity current.

Debris fields surrounding some islands, such as Hawaii, provide evidence that massive slides occur periodically when parts of oceanic ridge mountains collapse, carrying volcanic rocks far from the ridge. Such avalanches may cause intense turbidity currents and tsunamis (**Chap. 9**).

Accumulation Rates

All sediments are mixtures of particles from many different sources. The accumulation rate and type of sediment are determined by the relative quantities of particles from each source that are deposited at each location. For example, **Figure 6-15** schematically illustrates the accumulation of sediments at two hypothetical locations. The input rate of biogenous particles to the sediment is the same at each location, but the input rate of lithogenous sediment is much higher at the first location than at the second. The sediments that accumulate at Site A are predominant-

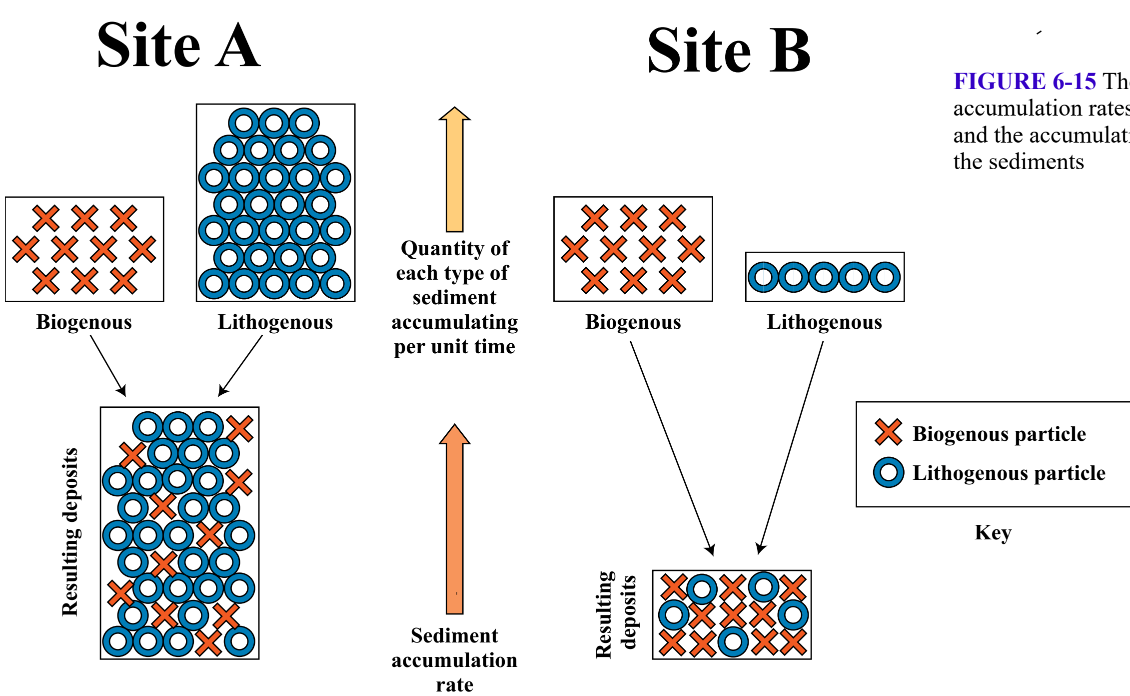


FIGURE 6-15 The relationships between the accumulation rates of different types of particles, and the accumulation rates and characteristics of the sediments

ly lithogenous, whereas the sediments at Site B are predominantly biogenous. However, the accumulation rate is much faster at A than at B. Because sediments are characterized by their predominant material, the Site A sediment would be called "lithogenous sediment," and the Site B sediment would be called "biogenous sediment," or **ooze**. An ooze is a sediment that contains more than 30% biogenous particles by volume. Where one type of organism is responsible for most of the biogenous particles, the ooze can be named after this type: pteropod ooze, radiolarian ooze, diatom ooze, foraminiferal ooze, and so on.

Lithogenous particles are the dominant input to ocean sediments. Most lithogenous material is discharged to the oceans from land as relatively large particles and deposited near river mouths and glaciers and in estuaries and wetlands. Sediment accumulation rates in these nearshore regions range from about 100 cm per 1000 years up to extreme rates such as the 7 m per year found in the delta of the Fraser River in British Columbia, Canada. Somewhat smaller but still large quantities of lithogenous sediment are transported offshore and are deposited on continental shelves. Such sediment can also reach the deep-ocean floor in areas where the continental shelves are narrow, or as turbidites. Many continental shelves are areas of high biological productivity (**Chap. 13**). Accordingly, sedimentation rates on the continental shelf and slope and within marginal seas, such as the Mediterranean, generally are about 10 to 100 cm per 1000 years.

In the deep oceans remote from land, lithogenous inputs are much reduced, and biogenous material, especially calcium carbonate, is dissolved before it can settle and be buried. Therefore, sedimentation rates in the deep-ocean basins are very low, approximately 0.1 cm per 1000 years. Under highly productive areas or on shallow seamounts or oceanic plateaus remote from land, the increased rate of sedimentation of biogenous material raises the sedimentation rate by about an order of magnitude, to approximately 1 cm per 1000 years.

Although sediments accumulate very slowly on a human timescale, they accumulate to substantial thickness in some places (**Fig. 6-16**). At a sedimentation rate of 0.5 cm per 1000 years, sediments approaching 1 km thick (850 m) can accumulate in 170 million years, which is the approximate age of the oldest (except for some remnants) oceanic crust.

CONTINENTAL MARGIN SEDIMENTS

Because most biogenous, hydrogenous, and cosmogenous sediments consist of silt- or clay-sized particles, they do not readily accumulate on continental shelves where currents are comparatively swift. Consequently, continental shelves are covered with lithogenous sediment of larger grain sizes, except in those areas where currents are slow or production of biogenous particles is very high.

On many continental shelves, lithogenous sediments accumulate continuously as new sediment is supplied by rivers and coastal erosion. The particles are sorted by grain size, and grain size generally decreases with distance from shore. However, on some continental shelves the supply of lithogenous material is very limited. For example, along the Mid- and North Atlantic coasts of the United States, most river-borne sediment and sediment eroded from the shore is trapped in estuaries and coastal lagoons. Consequently, little lithogenous sediment is transported offshore beyond the longshore drift system (**Chap. 11**). If the shelf is wide and currents are generally swift on the outer part of the shelf, little or no new sediment is supplied to the seafloor. Such areas of the continental shelf are not bare rock, but are covered by generally coarse-grained sediments, known as relict sediments, that were deposited under a different set of ocean conditions (**Fig. 6-17**).

Relict sediments, including those on the outer continental shelf of the U.S. east coast, contain terrestrial fossils and shells of organisms, such as oysters, that live only in water less than a few meters deep. These materials are too big to have been moved by currents but now are under 100 m or more of water. The reason

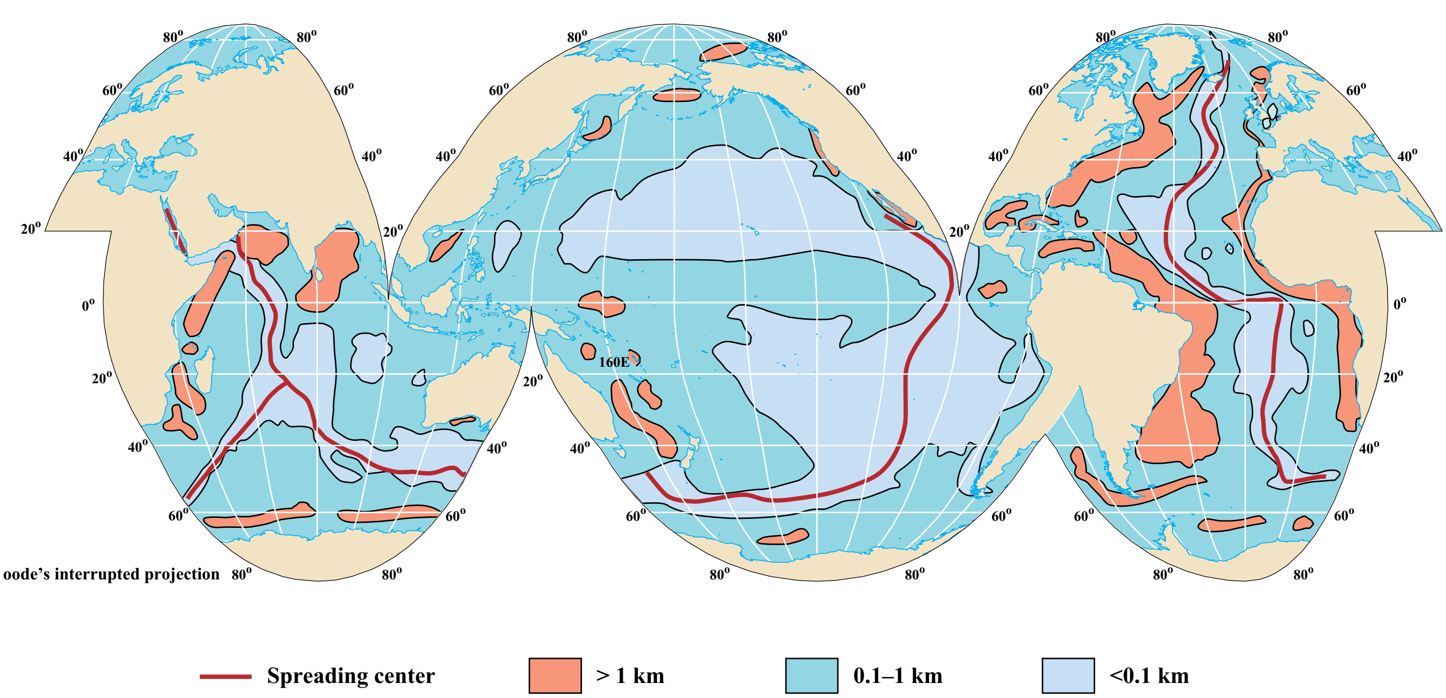


FIGURE 6-16 Thickness of accumulated sediments on the world's oceanic crust

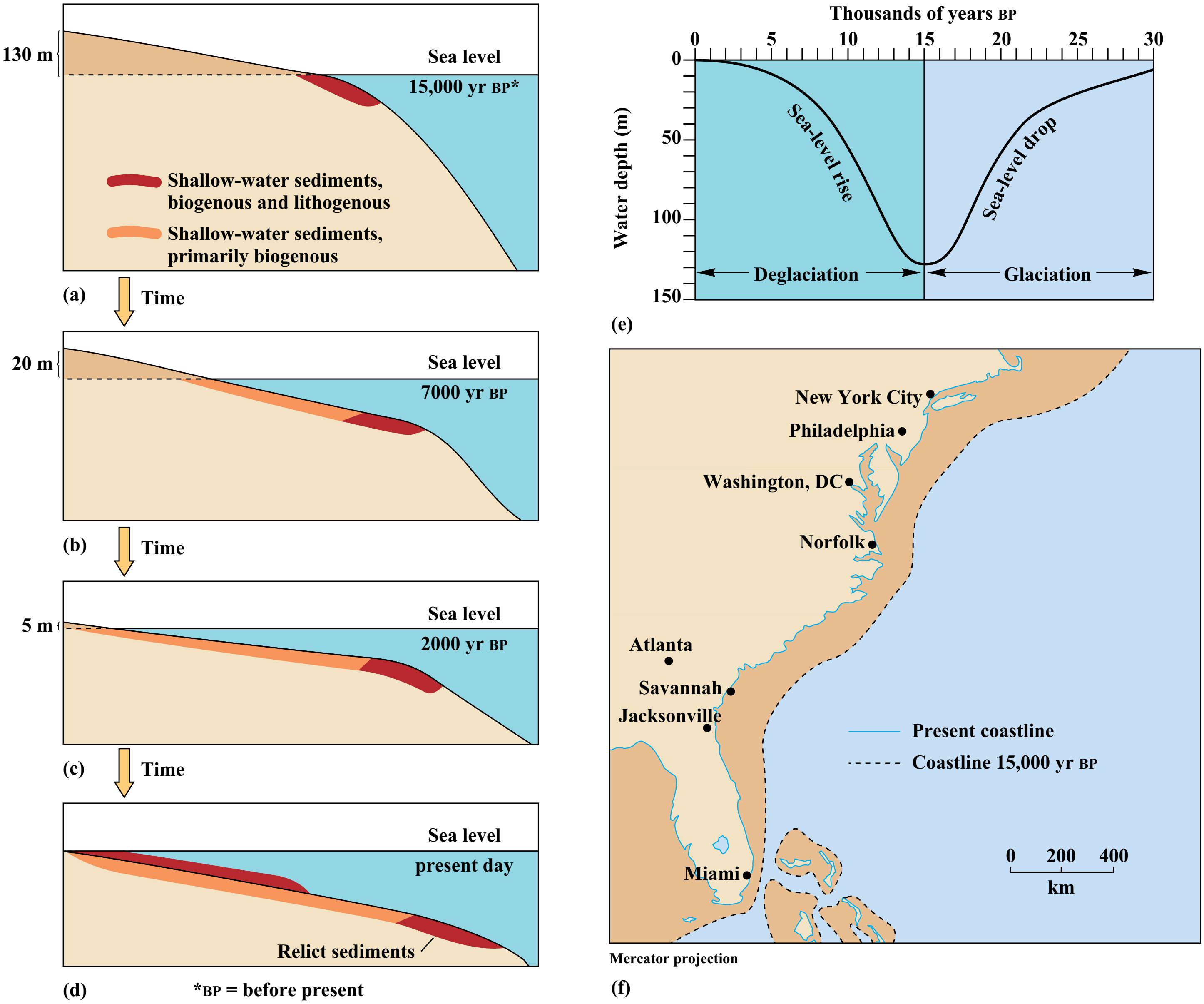


FIGURE 6-17 Relict sediments. (a) Sediments characteristic of shallow water were deposited on the outer continental shelf when sea level was lower. (b) As sea level rose, the relict sediments were not buried by more recent sediments, probably because of the rapid rise of sea level and the trapping of lithogenous sediments in newly submerged river valleys. (c) Shallow-water sediments formed close to the new shoreline as it migrated inland. (d) As sea level rose more slowly, the sediments close to the now slowly retreating shoreline were buried by terrigenous and shallow water biogenous sediments. (e) The history of sea-level change during the past 30,000 years shows the rapid rise between 19,000 and 4000 years ago and the much reduced rate of rise in the past 5000 years. (f) Locations of the Atlantic coastline of North America 15,000 years ago and today. Relict sediments are found primarily near the coastline location of 15,000 years ago.

is that the relict sediments were deposited several thousand years ago when the sea level was lower, and the coastline was near what is now the edge of the continental shelf.

The distribution and age of relict sediments are determined by the sea level. Much of the detailed history of recent sea-level change has been learned through studies of the location of relict sediments. Sea level has been rising from its most recent low point about 19,000 years ago to the present day, although not at a uniform rate.

DISTRIBUTION OF SURFACE SEDIMENTS

Surface sediments are the materials currently accumulating on top of older sediment. The older sediment buried below the surface layer may have a different character, if it was deposited under different conditions.

The distribution of pelagic surface sediments and their current rate of accumulation are shown in Figure 6-18- Continental margin sediments are not shown, because they are lithogenous except in areas dominated by coral reefs and in some areas of calcium carbonate precipitation. Each type of sediment and the factors that govern its occurrence are described briefly in the sections that follow.

Radiolarian Oozes

In the Pacific and Indian Oceans, under a region of high productivity that extends in a band across the deep oceans at the equator (Chap. 13), surface sediments are fine muds that consist primarily of radiolarian shells. Although both calcareous and siliceous organisms grow in abundance in this upwelling region, calcareous material dissolves and does not accumulate in sediments below the CCD. Radiolaria are prolific in tropical waters, and the rate of input of radiolarian shells to the sediment is much

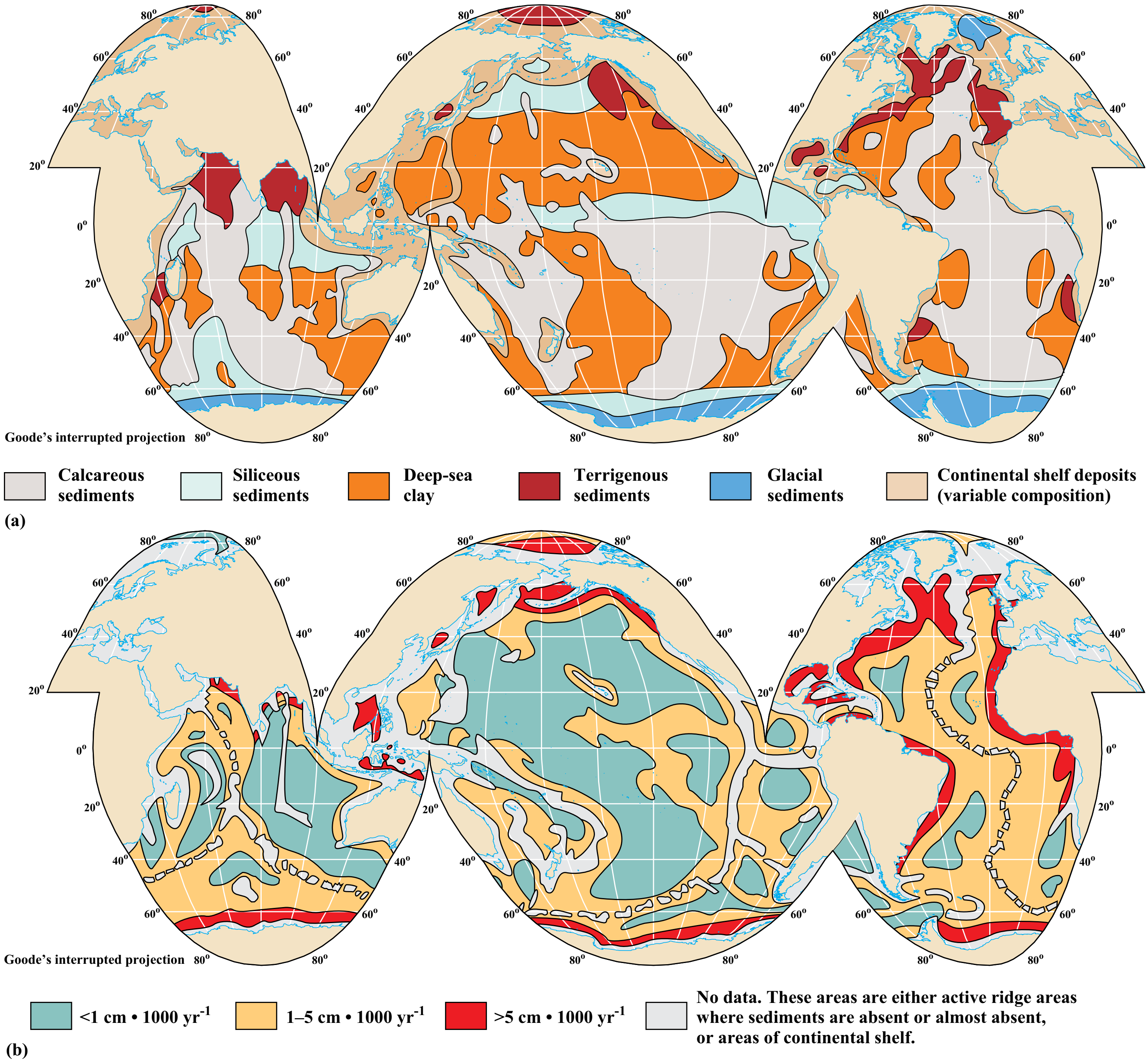


FIGURE 6-18 Present-day sedimentation. (a) The distribution of present-day surface sediments of the world oceans. (b) Approximate sedimentation rates in the present-day oceans. Note the higher rates of sedimentation near the continents. Surface sediments in these areas are predominantly terrigenous. Sediments are predominantly red clays in areas where the sedimentation rate is lowest. Note the generally slower rate of sediment accumulation on the abyssal floor of the Pacific Ocean compared to that of the Atlantic Ocean.

higher there than in other deep-sea mud areas. Because siliceous material dissolves slowly, many radiolarian shells are buried in the surface sediment before they fully dissolve.

The rate of input of fine-grained lithogenous particles that are resistant to dissolution is approximately the same in the equatorial band and in the adjacent areas where deep-sea mud accumulates. However, the rate of accumulation of radiolarian shells, and therefore the overall sedimentation rate, is much higher in the equatorial band. Therefore, the deep-sea lithogenous mud particles are "diluted" to become a minor component in radiolarian oozes.

The Atlantic Ocean has no band of radiolarian ooze because the accumulation rate of lithogenous sediment is much higher there than in the tropical Pacific and Indian oceans, the seafloor is relatively shallow, and the CCD is relatively deep. The radiolarian content of tropical Atlantic sediments is diluted and masked by the greater contributions of calcareous and lithogenous sediment. Hence, the overall sedimentation rate in the tropical Atlantic Ocean exceeds the sedimentation rate in the tropical Pacific and Indian oceans.

Diatom Oozes

Diatoms dominate the siliceous phytoplankton in upwelling areas except in the tropical upwelling zone where radiolaria dominate. Upwelling and abundant diatom growth occur in a broad band around Antarctica and in the coastal oceans along the west coasts of continents (eastern ocean margins) in subtropical latitudes (**Chaps. 8, 13**). Diatom oozes dominate in these locations if the sedimentation rates of lithogenous and calcareous particles are relatively slow.

Aside from large coastal inputs from Antarctica's glaciers, sedimentation in the Southern Ocean is limited because of the paucity of landmasses and river inflows. Hence, the accumulation rate of lithogenous particles is slow in the band of high diatom productivity that surrounds Antarctica, and the sediments are dominated by diatom frustules.

Terrigenous sediment inputs are also limited on the west coasts of North and South America because of the limited drainage areas of coastal rivers and the trapping of terrigenous sediment in the trenches that parallel these coasts. Because the seafloor is deep in the Pacific coast upwelling regions and the CCD is relatively shallow in the Pacific Ocean, calcareous particles are dissolved and do not accumulate in sediments. Diatom frustules therefore dominate the sediments in the north and south subtropical Pacific offshore basins. In contrast, on the eastern ocean boundaries in the Atlantic Ocean (southern Europe and West Africa) and Indian Ocean (Australia), in areas where the seafloor is relatively shallow, calcareous particles accumulate faster than diatomaceous particles, and the sediments are calcareous.

Diatoms are very abundant in Northern Hemisphere cold-water regions (Bering Sea and North Atlantic), just as they are near Antarctica. However, lithogenous sediment inputs are greater in the Northern Hemisphere than in the Southern Ocean, and much of the seafloor is above the CCD. Therefore, diatomaceous particles are diluted by calcareous and lithogenous particles, and diatomaceous oozes accumulate only in limited deep areas of the Bering Sea and northwestern Pacific Ocean.

Calcareous Sediments

In areas where the seafloor is shallower than the CCD, calcareous sediment particles accumulate faster than deep-sea clays. Therefore, where the seafloor is shallower than the CCD and terrigenous inputs are limited, calcareous particles are a major component of the sediment. Calcareous sediments are present on oceanic plateaus and seamounts and on the flanks of oceanic ridges.

Deep-Sea Clays

Few large particles are transported into the deep oceans remote from land unless they are carried by turbidity currents (mostly in the Atlantic Ocean). Most deep-ocean areas far from land sustain relatively poor productivity of marine life (**Chap. 13**). Because of the great depth, the relatively small quantities of biogenous material that fall toward the seafloor are mostly dissolved before reaching, or being buried in, the sediment. This is especially true for calcareous organisms in areas where the seafloor is below the carbonate compensation depth (CCD). Therefore, the particles deposited on the deep-sea floor far from land are mainly fine lithogenous quartz grains and clay minerals.

Deep-sea clays (sometimes called "red clays") are reddish or brownish sediments that consist predominantly of very fine-grained lithogenous material. Everywhere but at high latitudes and in a high productivity band extending across the equatorial Pacific and Indian Oceans, they cover deep-ocean floor that is remote from land. The reddish or brownish color is due to oxidation of iron to form red iron oxide, which we know as rust, during the slow descent of the particles to the seafloor. Oxidation continues on the seafloor until the particles are buried several millimeters deep. The sedimentation rate is very low, so burial is very slow and the interval during which oxidation occurs is very long.

Siliceous Red Clay Sediments

In the deep basins of the North and South Pacific, South Atlantic, and southern Indian Oceans, transitional areas are present between the central deep basin, whose surface sediments are red clays, and higher latitudes, where surface sediments are diatom oozes. Sediments in these transitional areas are mixtures of deep-sea clay and diatomaceous sediments in varying proportions. The transitional sediments are often classified with deep-sea clays and often called "deep-sea muds."

Ice-Rafted Sediments

In the Arctic Ocean, northern Bering Sea, and a band immediately surrounding Antarctica, the sediment consists largely of lithogenous material carried to the oceans by glaciers. It includes sand, pebbles, and even boulders at distances of up to several hundred kilometers offshore from the glaciers. Some lithogenous material is released directly to the ocean when the edge of the ice shelf or glacier melts. Some is transported by icebergs, also called "ice rafts," that break off the glaciers. The material released and deposited as the ice melts is known as glacial, or "ice-rafted," debris. Because glaciers can carry huge quantities of eroded rock, the sedimentation rate can be very high. Biogenous particles therefore are diluted by and mixed with much larger volumes of ice-rafted lithogenous material.

Terrigenous Sediments

Sediments dominated by terrigenous particles are present near the mouths of major rivers and may extend hundreds of kilometers offshore. Such sediments are deposited in the northern Arabian Sea, the Bay of Bengal, the Gulf of Mexico, and the Atlantic Ocean near the mouth of the Amazon River in Brazil. Terrigenous sediments transported by turbidity currents are present at the foot of the continental slope in the North Atlantic. On the North Pacific coast of North America, rivers carry large sediment loads from the glaciers in the mountains of British Columbia and Alaska. Many of the glaciers terminate and melt before reaching the ocean. Ice rafts are few, so coarse sand, pebbles, and boulders are not transported beyond the immediate vicinity of the glacier termination. However, large quantities of glacially ground sand, silt, and mud are transported to the ocean by rivers, and offshore by waves and currents.

Hydrothermal Sediments

The central basin of the Red Sea is the only area known to have hydrothermal mineral deposits as the dominant sediment type. The other locations where such sediments accumulate are small and scattered throughout the oceans, mostly on the oceanic ridges. The extent of such deposits is not known, because only a very tiny fraction of the oceanic ridge system has been surveyed by methods that reveal hydrothermal vents and their associated sediments. Vents have been found on oceanic ridges in all oceans and are likely to be present at intervals of a few kilometers to hundreds of kilometers along the entire oceanic ridge system.

THE SEDIMENT HISTORICAL RECORD

Sediments accumulate continuously by deposition of new layers of particles on top of previously deposited layers. Therefore, individual layers that make up the sediment below the sediment surface were deposited progressively earlier as depth below the sediment surface increases. The type of particles deposited at a specific location and time depends on many factors, such as the proximity to land, whether rivers or glaciers flowed into the adjacent coastal ocean, the type and amount of dust in the atmosphere, and the composition and production rates of organisms in the water column. These factors do not remain constant over geological time, because they are affected by plate tectonic movements, climate changes, and changes in volcanic activity, all of which may be related to each other in complex ways. Such changes, singly or in combination, can move coastlines, change seafloor depth, raise mountains, alter rivers, create and melt glaciers, change atmospheric dust composition by creating or reducing deserts or by altering the amounts of volcanic material injected into the atmosphere, and change the temperature, salinity, and nutrient distributions in the oceans. Each of these changes affects the type of sediment that accumulates. Hence, buried sediments may be very different from those currently accumulating.

Buried sediments provide information about the conditions at the time they were buried. Ocean sediments therefore can provide a history of plate tectonics and climate changes covering about 170 million years, the approximate age of the oldest remaining oceanic crust in the major oceans. Older fragments of oceanic crust are known to exist but it is not yet known whether they support undisturbed sediment layers. To read the history of ocean sediment layers, we must be able to determine the date at which a particular layer of sediment was deposited. Then we must examine the composition of the sediment particles to reconstruct the sedimentation conditions at the time the layer was formed.

Stratigraphy is the study of the Earth's history through investigation of the sediment layers beneath the ocean floor (and on land where sediments have been compressed and converted into sedimentary rock). Stratigraphic studies of ocean sediments usually involve the examination of sediment cores (**Chap. 3**), which are sectioned into slices 1 or 2 cm thick (**Fig. 6-19**). Each slice represents the accumulation during a period of thousands or tens of thousands of years, depending on the sedimentation rate.

The stratigraphic record can be complicated to read. Gaps may appear in the record if previously deposited sediment was eroded because of a temporary increase in current speed. Furthermore, adjacent layers may have had substantially different sedimentation rates. Biological activity also may have altered the historical record. Bottom-dwelling organisms rework, or mix, the upper few centimeters or tens of centimeters of sediment as they consume the organic matter, feed on other sediment dwellers, or build burrows. Their churning of the sediments is called bioturbation. Despite these limitations, the stratigraphic record provides vital information about the Earth's history, including previous climate changes. Understanding climate changes will enable us to better understand the factors that control changes in climate such as those expected to result from enhancement of the greenhouse effect.

Sediment Age Dating

A fundamental requirement of stratigraphic studies is that the age (date of deposition) of the sediment layers be determined. Methods for determining sediment ages include radioactive dating (**CC7**), magnetic dating, and biological dating.

Radioactive dating of sediments is difficult because often the assumptions on which the technique is based are not met. Therefore, researchers commonly use several different methods to be certain of a measured age, including dating of sediments based on their magnetic properties. To explain how magnetic dating is done, we need to understand first that the magnetic poles appear to have been close to the Earth's north and south poles of rotation for billions of years, even though they wander slightly. However, the Earth's magnetic field completely reverses periodically for reasons we do not yet fully understand. During a reversal, the north magnetic pole is located close to the geographic South Pole and the south magnetic pole close to the geographic North Pole. The sequence of these reversals is well documented from studies of terrestrial rocks. In sediments, particles that contain iron or other magnetic materials tend to align with the magnetic field during or soon after deposition in sediments. As a result, the magnetic properties of sediment vary with depth below the sediment surface, changing at each depth in the sediment column that corresponds to a magnetic reversal. Measuring the magnetic properties of the sediment layers and correlating the depth of layers where changes in magnetic field occur with the known dates of magnetic pole reversals therefore allows us to date the sediments.

Another important dating technique is based on the fossil species found within the sediments. Many of the marine species whose remains make up the biogenous fraction of older sediments have become extinct, have been replaced by new species, or have evolved into

FIGURE 6-19 Sediment cores, such as this deep-sea drilling core, are first cut down the middle and then segmented into small slices, each representing the accumulation of sediments during a period of hundreds or thousands of years in the past.

substantially different forms. Consequently, the ages of sediments at different locations often can be matched by comparison of the species compositions of organisms found in their biogenous fractions. If the sediment at a specific depth at one sampling site contains the fossil remains of all the same species that the sediment at a different depth at another sampling site contains, the sediments are likely to be the same age. Thus, we can date sediments in relation to each other and determine relative sedimentation rates.

Diagenesis

Interpreting the stratigraphic record requires an understanding of **diagenesis**, the physical and chemical changes that occur within sediment after it is buried. In some cases, during the millions of years they lie buried, minerals are altered from one form to another as weathering processes that began on land or in suspended sediment continue. However, the more important diagenetic changes occur in the pore water (interstitial water) that is trapped between the sediment grains.

Chemicals that dissolve in pore waters during diagenesis include silica and calcium carbonate, which continue to dissolve from buried siliceous and calcareous material. Also during diagenesis, other substances are released by continuing oxidation of sediment organic matter, and these too may be dissolved in the pore water. Chemicals dissolved from the sediment particles during diagenesis migrate upward by **diffusion** toward (and in some cases into) the overlying water. As sediments accumulate, their weight compacts the underlying sediment. In the process, pore waters are slowly squeezed out and migrate upward through the sediment.

As organic matter decomposes in the sediment, dissolved oxygen in the pore water is consumed and oxidation continues with oxygen from nitrates and then sulfates (**Chap. 12**). The sulfates are reduced to sulfide, and the presence of sulfide dramatically changes the solubility of many elements. Iron, manganese, and several other metals form sulfides that are much more soluble than their hydrated oxides, which form in oxygenated water. In sediment that has enough organic matter to produce sulfides, iron and manganese are dissolved in pore waters that migrate upward. The metal sulfides migrate with the pore water until it approaches the sediment surface, where it encounters dissolved oxygen diffusing down into the sediment or introduced by bioturbation. The iron and manganese then are oxidized and redeposited on sediment grains. As sediment accumulates, these elements may be dissolved and moved upward continuously, accumulating in a layer near the sediment surface where sulfide and oxygen mix.

In contrast to iron and manganese, the sulfides of many elements (e.g., copper, zinc, and silver) are less soluble than their hydrated oxides. These metals are not dissolved in pore waters that contain sulfide, so they do not migrate up toward the oxygenated surface sediments and overlying seawater. On the contrary, these elements can diffuse with oxygenated seawater downward into the sediment, where they can be converted to sulfides and buried with the accumulating sediment. The complex migrations of elements within pore waters and the associated changes in the nature of the sediment particles must be understood if we are to deduce the nature of the sediment particles at the time of their deposition and thus read the stratigraphic record.

Diagenetic processes are also important to life. For example, the nutrients nitrate, phosphate, and silicate are transported to the sediments in detritus that falls to the sediment surface. The nutrients are released to the pore water as organic matter is oxidized and they then diffuse back into the water column, where they can be reused by living organisms.

Tectonic History in the Sediments

Much of the history of tectonic processes can be revealed by studies of the changes in sediment characteristics with depth below the sediment surface. For example, consider the sediment layers that have accumulated at locations between the oceanic ridges and continents (**Fig. 6-10**). Remember that the seafloor is progressively older with increased distance from the oceanic ridge (**Chap. 4**). However, the older seafloor was once at the oceanic ridge and sank isostatically (**CC2**) as it cooled, so it becomes progressively deeper. Remember also that all sediments are mixtures (**Fig. 6-15**). As a result, sediment deposits can reveal the history of plate tectonic movements. For example, the age of a buried calcareous sediment layer can reveal the times during which the seafloor was at depths shallower than the CCD.

In reality, reading the sediment history is complicated because important factors, such as the CCD level, have undoubtedly varied over the millennia. In addition, the locations at which certain types of organisms are abundant may have changed. Currently, siliceous diatoms abound near the poles, and siliceous radiolaria dominate near the equator. However, the distributions of these organisms might have been very different at times in history when the climate was different from the present-day climate.

Climate History in the Sediments

The past 170 million years of the Earth's climate history are preserved in ocean sediments, primarily in biogenous particles. Each marine species that contributes calcareous and siliceous material to ocean sediments has an optimal set of conditions for growth. Although nutrient concentrations, light intensity, and other factors are important, temperature is crucial for most species. Hence, fossils in sediments can be used to determine the geographic variations in ocean water temperatures.

Because different species of foraminifera thrive at different temperatures, the species composition of foraminiferal fossils in buried oozes can be used to study past climate. However, because some species have evolved or become extinct, the temperature tolerances of fossil species often are assumed to be the same as those of modern species. This assumption may not be correct.

Past climate information also can be obtained from measurements of the ratio of oxygen **isotope** concentrations in calcareous sediments. The basis for this method is somewhat complicated. Oxygen has three isotopes: oxygen-16, oxygen-17, and oxygen-18. Almost all ocean water molecules consist of either an oxygen-16 atom combined with two hydrogen atoms, or an oxygen-18 atom combined with two hydrogen atoms (the concentration of oxygen-17 is very small). Oxygen-16 is lighter than oxygen-18. Consequently, water containing oxygen-16 is lighter than water containing oxygen-18. The lighter water molecules evaporate more easily than the heavier molecules. Hence, when seawater evaporates, the water vapor is enriched in oxygen-16. When the Earth's climate cools, more water is precipitated and stored in the expanding glaciers and polar ice sheets. Because this water was evaporated from the oceans, oxygen-16 is transferred preferentially to the ice, the ratio of oxygen-16 to oxygen-18 in ocean water decreases, and sea level falls. Organisms that live in seawater incorporate the oxygen-16 and oxygen-18 of seawater into their calcium carbonate body parts in a ratio that is determined partly by the ratio of these two isotopes in the seawater. Therefore, the ratio of oxygen-16 to oxygen-18 in calcareous sediment can be used to deduce climatic characteristics and sea level at the time of deposition. The ratio is high when the climate is warm and sea level is high, and low when sea level is lowered by evaporation and more water is stored on land as ice and snow.

In the same way that water molecules containing oxygen-16 and oxygen-18 evaporate at slightly different rates, molecules containing these oxygen isotopes react at slightly different rates in the chemical processes that produce the calcium carbonate of calcareous organisms. The ratio of oxygen-16 to oxygen-18 in the calcareous parts differs between species, but for a single species it depends on the ratio in the seawater and the seawater temperature. At any given time, the oxygen isotope ratio of ocean surface water is relatively uniform throughout the oceans. Hence, calcareous remains of the same species deposited in different parts of the ocean at the same time have different ratios of oxygen-16 to oxygen-18 because the water temperatures were different where the particles were deposited. A low ratio of oxygen-16 to oxygen-18 indicates colder water.

Studies of sediments aimed at revealing the Earth's climate history have become more important because of growing concern about human enhancement of the greenhouse effect. An understanding of past climates can help us understand future climate change due to both natural causes and human activities. Stratigraphic studies are becoming ever more sophisticated and are now beginning to reveal details of ancient ocean current systems and ocean chemistry, including carbon dioxide concentrations. Such information is continually improving our understanding of the complicated feedback mechanisms between the atmosphere and oceans (**Chap. 7**, **CC9**).

History of Ocean Acidification and Deoxygenation in the Sediments

As we have seen in the previous section, ocean sediments can be used to study the variation of temperatures in Earth's past. Sediments also provide a historical record of changes in ocean water chemistry. For example, past changes in the acidity of the oceans and of the extent of oxygenation of the oceans can be revealed by studies of sediments and the fossils found in sediments. The techniques used to extract acidity and oxygenation information from sediments are complex and beyond the scope of this text. However, the results of such studies provide important information on the likely effects of anthropogenic emissions on ocean ecosystems. This information suggests that acidification and **deoxygenation** may have severe adverse effects comparable to, and perhaps greater than, the adverse effects of climate change.

Evidence of Mass Extinctions

Earth has experienced at least five major mass extinctions of species. All but one of these occurred farther in the past than the record provided by current ocean sediments. These earlier extinctions can be studied by records left in sedimentary rocks that were formed from sediments present in the oceans when each event occurred. Sedimentary rocks are more difficult to study than sediments since they are more likely to have been altered by chemical and physical processes that occurred after they were formed and uplifted by tectonic processes. However, each of the mass extinctions has been investigated and it has been learned that, in at least two of these events, the extinction was the result of climate warming due to rising atmospheric carbon dioxide levels, and associated ocean acidification and deoxygenation. The atmospheric carbon dioxide rise in most or all extinctions was likely due to increased volcanic activity. Volcanic activity, can introduce large amounts of dust and gas to the atmosphere, altering global climate and causing the extinctions. For example, the 1883 eruption of Krakatau, a relatively small eruption in comparison with some that have occurred in the past, is known to have affected the world climate for a decade. Large increases in volcanic activity are known to have occurred periodically in Earth's history

The rate of increase of carbon dioxide in the atmosphere during Earth's mass extinctions is much slower than that Earth is now experiencing due to anthropogenic releases but past extinctions were not instantaneous events and instead take place over at least thousands of years. Volcanic activity can release large quantities of sulfur that have a similar but different effect on the atmosphere and oceans than carbon dioxide. Despite these differences, a better understanding of past extinctions can provide us with a better understanding of the relative effects of anthropogenic climate change, ocean acidification, deoxygenation in the future and of whether or not anthropogenic inputs may lead to another mass extinction.

Investigating a Mass Extinction

About 65 million years ago, the last of the dinosaurs became extinct. Fossil and sediment records show that, at the same time, about 70% of all species then on earth including more than half of all species of marine animals became extinct. Although other extinctions have been found at different times in the sedimentary rock record, the dinosaur extinction was among the most dramatic.

For a number of years, there was a major debate about what caused this extinction. It was known that a major increase in volcanic activity occurred at about the time of the extinction event. However, dating of events such as the eruption and extinction is

FIGURE 6-20 The location of the Chicxulub crater, which is thought to be the impact site of a massive meteorite that may have been responsible for climate changes that, in turn, led to the extinction of the dinosaurs. The crater is at least 180 km in diameter, but it may be twice that size.

difficult and until recently the volcanic event, called the Deccan Traps, flood basalt eruption in China was thought to have occurred about 300,000 years before the extinction event. Instead, evidence supported the hypothesis that a meteorite impact near Chicxulub, Mexico could have been responsible for the extinction. The studies leading to this meteorite impact conclusion provide a good illustration of how past events can be deduced from sediments.

The area that is now the Yucatán Peninsula of Mexico was underwater 65 million years ago at a depth of about 500 m. Magnetic surveys in this area reveal a circular feature (**Fig. 6-20**), 180 km in diameter, that may be the buried remains of the impact crater of a huge meteorite. Studies of sedimentary deposits of the appropriate age from many areas surrounding this circular feature, now called Chicxulub, show an unusual series of layers at the level that corresponds to deposition 65 million years ago.

Sediments below the strange layers are fine-grained biogenous oozes of the type normally deposited at a depth of approximately 500 m. The unusual layers immediately above these normal sediments have characteristics that could be explained by the sequence of events that followed a massive meteorite impact. The deepest and oldest strange layer consists primarily of rounded, coarse grains several millimeters in diameter. These grains were not rounded by water weathering. They are glassy tektites, which form when rocks are melted by meteorite impacts, ejected into the atmosphere where they resolidify, and fall back to the Earth. This layer also contains high concentrations of iridium, an element rare on earth but more common in meteorites. An iridium rich layer is found in 65-million-year-old sediments from all parts of the oceans but decreasing in concentration with distance from Chicxulub, providing evidence that a major meteorite hit had spread dust through the entire earth's atmosphere at this time and that the impact took place near Chicxulub.

The next higher layer consists of coarse-grained sediments that contain fossilized wood and pinecone fragments. These materials are not normally found in marine sediments. It is hypothesized that this layer was created when the meteorite impact generated a huge tsunami (**Chap. 9**) that smashed onto land, tore rocks and trees loose, and carried them back to sea as it receded. The tsunami may have been so large that it sloshed back and forth across the Gulf of Mexico and around the world for at least several days. It must have been many times higher than the tsunamis that hit Indonesia in 2004 and Japan in 2011, perhaps several hundred meters or more high.

Above the two older extraordinary layers is a series of layers of progressively finer-grained sand mixed with other mineral particles that are not normally found in marine sediments. These progressively finer-grained layers may consist mostly of beach sands and eroded soil particles suspended and transported to the deep water by waves generated by the meteorite impact. As wave energy decreased after the initial meteorite impact and tsunami, progressively smaller particles would have been deposited in the same way that turbidites are. Much of the finer-grained material overlying the graded sequence may have come from the fallout of dust ejected into the atmosphere during the impact.

Above these finer-grained layers, the sediments finally grade into normal biogenous oozes similar to the ones that underlie the unusual layers but with fewer and different species. For example,. below the anomalous layers foraminifera species were large, abundant and diverse, but above they were almost absent, smaller and of many fewer species.

There are many possible variations of the details of the events that deposited the anomalous sediment layers 65 million years ago. For example, one suggestion is that the tsunami may not have been caused directly by the impact of the meteorite on the ocean. Instead, the impact may have caused a magnitude-11 earthquake (about a million times more powerful than the 1989 Loma Prieta earthquake near San Francisco), which could have sent giant turbidity currents down the continental slope and thus generated massive tsunamis.

Additional details of the Chicxulub impact have emerged as studies of the impact continued. For example, it has been estimated that the object that hit the Earth was about 10 km in diameter, that the impact created a crater more than 180 km in diameter and that it must have caused strong earthquakes, and most likely a massive fireball. The meteorite impacted carbonate and sulfur rich rocks and it has been estimated that 100-150 gigatons of sulfur were released to the atmosphere. This would have caused devastating acid rains while the dust blown into Earth's upper atmosphere would have blocked enough sunlight to substantially cool Earth's climate for decades or longer and, perhaps more importantly, to drastically reduce the sunlight available for photosynthesis. Photosynthetic organisms are the base of both terrestrial and marine food chains, so the loss of available food would have caused many species populations to crash to, or close to, extinction even if they were not adversely affected by the initial firestorm, earthquakes and tsunami or the subsequent acid rain.

As we can see, the odd sediment layers in the Gulf of Mexico have begun to reveal a detailed picture of what must have been one of the largest events in Earth's history. This evidence appeared to support a hypothesis that the mass extinctions was caused by the meteorite impact rather than by a major increase in volcanic activity. However, as often happens in science, the simple answer is not always correct. More recent studies of fossils in sediments from before and after the impact, and better dating of sediment layers have now revealed the impact is likely not the single cause of the mass extinction. This more recent data now suggests that this massive volcanic eruption took place over a time period that straddles the extinction with at least four large pulses of volcanic activity one of which preceded the extinction. Also, more detailed dating of the disappearance of species in the fossil record showed that a number of species became extinct before the impact while a comparable number of species became extinct after the impact.

It now appears that this mass extinction, that included an end to the dinosaurs and led to the rise of mammals, was caused by a double disaster. First, there was an episode of increased volcanic activity on the planet causing rapid climate warming and almost certainly acidification and deoxygenation (as scientific evidence supports likely occurred during other extinctions). Second, during the period of increased volcanic activity there was the massive Chixulub meteorite impact.

The sediments related to this mass extinction will undoubtedly be studied extensively, and their meaning debated for many years and this double disaster scenario may change again, but there are two important lessons to learn. First, environmental events do not always have a simple single cause. They are most often a combination of several processes interacting with each other. Second, our tendency to think of natural events on human time scales can be misleading. A mass extinction that takes

100,000 years or longer to take place is a very abrupt change on the timescale that controls Earth's climate and life. We think of the anthropogenic impacts on Earth as having taken place over a long period of centuries, but in geological time humans have caused changes that are far faster than anything seen on Earth before, or at least since the Earth's formation and early history.

Evidence of Other Impacts

Researchers believe that there may have been a number of other large impacts similar to the Chicxulub event at different times in the past, and they are searching for evidence of the impact craters. A number of locations have been identified as possible impact sites, including an area off the northwestern coast of Australia, where there may have been an impact about 250 million years ago, and an area at the mouth of the Chesapeake Bay, where an impact may have left an 85-km-wide crater about 35 million years ago. Decades of careful research will undoubtedly be needed to demonstrate that meteorite impacts did or did not occur in these areas and others. If confirmed, additional research will be needed to investigate and identify the effects that those impacts may have had on the biosphere.

CHAPTER SUMMARY

Classification of Sediments.

Sediments can be classified by their predominant grain size or by the origin of the majority of their particles. Grain size ranges usually are classified as gravel, sand, silt, and clay, in order of decreasing size.

Lithogenous Sediments.

Lithogenous sediment particles are primarily chemically resistant minerals produced by weathering of continental rocks. They are transported to the oceans by rivers, glaciers, waves, and winds. Lithogenous sediments are deposited in thick layers beyond the mouths of some rivers, but many rivers trap sediments in their estuaries. Fine-grained dust, especially from deserts, can be transported by winds for thousands of kilometers before settling on the sea surface and then sinking to the seafloor. Some volcanic eruptions eject large quantities of ash that reach ocean sediment in a similar way.

Biogenous Sediments.

Biogenous sediments are the remains of marine organisms. In most cases, the organic matter is decomposed before the hard inorganic parts are deposited and buried.

Some marine species have calcareous or siliceous hard parts, both of which dissolve in seawater. Most siliceous diatom frustules dissolve very slowly and become included in sediments. The intricate siliceous shells of radiolaria generally dissolve before being buried in sediments in all but tropical waters where radiolaria are abundant. Because the solubility of calcium carbonate in seawater increases with pressure, the rate of dissolution of calcareous hard parts increases with depth. Calcareous shell material has two forms: calcite and aragonite. The aragonite of pteropods is dissolved much more easily than the calcite of foraminifera. Pteropod oozes are present only in relatively shallow water, whereas foraminiferal oozes predominate at intermediate depths. Below the carbonate compensation depth, little or no calcareous material survives dissolution to be incorporated in the sediment. The CCD varies between oceans because of differences in dissolved carbon dioxide concentrations, and it also varies historically with climate changes.

Hydrogenous Sediments.

Hydrogenous sediments, which are less abundant than terrigenous (lithogenous) or biogenous sediments, are deposited by the precipitation of minerals from seawater. They include manganese nodules that lie on the sediments in areas of the deep ocean, phosphorite nodules and crusts on continental shelves beneath oxygen-deficient water, calcium carbonate deposits in a few shallow areas where water temperature is high and dissolved carbon dioxide concentration is low, and evaporites (salt deposits) in coastal waters with limited water exchange with the open ocean, low rainfall, and high evaporation rate. Hydrogenous sediments also include metal-rich sediments accumulated by the precipitation of minerals from water discharged by hydrothermal vents.

Cosmogenous Sediments.

Cosmogenous sediment particles are fragments of meteorites or tektites created by meteorite impacts. They constitute only a tiny fraction of the sediments. Many are found as cosmic spherules.

Sediment Transport and Deposition.

Large particles are deposited quickly, unless current speeds are high, and they are difficult to resuspend. Large particles collect close to river mouths, glaciers, and wave-eroded shores. Because orbital velocity in waves is higher than ocean current speeds, large particles can be transported in the nearshore zone but cannot be transported far from shore. Smaller particles may be carried long distances before being deposited.

Turbidity currents that resemble avalanches carry large quantities of sediment down continental slopes and onto the abyssal plains. Graded beds of turbidites are present in deep-ocean sediments, and they are common on the abyssal floor near the continents, except where there are trenches.

Sediment Accumulation Rates.

Sediments are mixtures of particles of different origins. Sediment characteristics are determined by the relative accumulation rate of each type of particle. Sediment accumulation rates range from about 0.1 cm per 1000 years in the deep oceans to more than 1 m per year near the mouths of some rivers.

Continental Margin Sediments.

Because fine-grained sediments are transported off the continental shelf by currents, shelf sediments are generally sand- and silt-sized lithogenous particles. Relict sediments, deposited when sea level was lower and containing remains of terrestrial and shallow-water organisms, are present in some areas on the continental shelf.

Distribution of Surface Sediments.

Deep-sea clays are present far from the continents on the abyssal plains. Radiolarian oozes occupy a band of high primary productivity that follows the equator, but not in the Atlantic, where the radiolarian shells are diluted with lithogenous sediment. Diatom oozes are present in high-latitude areas and in other upwelling areas where primary productivity is high and lithogenous inputs are low. Sediments dominated by terrigenous particles are deposited near the mouths of major rivers. Calcareous sediments dominate where the seafloor is shallower than the CCD and lithogenous inputs are low. Hydrothermal sediments are deposited in some areas on oceanic ridges.

The Sediment Historical Record.

Sediments accumulate continuously and provide a record of the Earth's history. The stratigraphic record can be complicated because sediment layers may be disturbed or chemically and physically altered by diagenesis. Stratigraphic studies of ocean sediments have revealed much of the Earth's climate and tectonic history during the past 170 million years. For example, such studies have provided information about the impact of a large meteor that may have contributed to the extinction of the dinosaurs and other species 65 million years ago.

KEY TERMS

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

abyssal fan (plain)
adsorption
biogenous
biogeochemical cycle
bioturbation
calcareous
carbonate compensation depth (CCD)
coastal plain
coccolithophores
cohesive
colloidal
continental shelf (margin, slope)
cosmogenous
decomposer
delta
deposition
detritus
diagenesis
diatom
diffusion
electrostatic
erosion
evaporite
excreted
fecal pellets
fjord
foraminifera
fossil
frustule
glacier (glaciation)
graded bed
grain size
hard parts
hydrogenous
hydrothermal vent
lagoon
leaching
limestone
lithogenous
manganese nodules
marginal sea
mollusk
ooze
pelagic
phytoplankton
pore water
pteropod
radiolaria
relict sediment
residence time
resuspended
runoff
sedimentation rate
shore (shoreline)
siliceous
slump
sorted
steady state
submarine canyon
suspended sediment
terrigenous
turbidite
turbidity current
turbulence
weathered
wetland
zooplankton

STUDY QUESTIONS

1. Sediments are deposits of particles of many different sizes from many different sources. Why can we classify sediments by their grain size?
2. List the four principal types of sedimentary particles classified by their origin. Can we classify sediments from all locations according to these four groups? Explain why or why not.
3. A sample of the surface sediments on the continental shelf a kilometer offshore contains large amounts of organic matter compared to other continental shelf sediments. What can you conclude or hypothesize about the characteristics of this region?
4. What conditions are necessary for deposits of metal rich sediments to be formed at hydrothermal vents?
5. Why would you not expect to find turbidites near the East Pacific Rise?
6. Why do we find relict sediments that have only a small amount of overlying sediment on the continental shelf off the northeastern United States, but not off the coast of Louisiana?
7. Describe the characteristics of surface sediments on the deep-sea floor. Why do they have these characteristics and how do they vary with location?
8. If you drilled into the sediments of the deep-sea floor just beyond the continental rise of the northeastern United States, what layers of sediment would you be likely to encounter with depth?
9. A 30-cm-long sediment core has a high concentration of organic matter throughout its length, but oxygen is not depleted in the pore waters. What might explain this?

CRITICAL THINKING QUESTIONS

1. Fifty centimeters deep in a 10-m-long sediment core taken from the center of an abyssal plain in the Pacific Ocean, a thin layer of sediment is found that has a much larger grain size than the sediment either below or above it. (a) Without knowing anything more about the material, give the possible explanations for this layer. (b) How could you determine which one of these alternative explanations was most likely?
2. If you were dating different levels within a sediment core and found a layer that was older than both the layer above it and the layer below it, how would you explain this?
3. Offshore oil drilling now takes place in very deep waters of the continental slope and may eventually take place at abyssal depths. In many cases, plans call for oil and gas produced from these wells to be collected through short pipelines from several wells in a small area and then shipped ashore on tankers. (a) Why might it not be a good idea in some deep locations to build pipelines on the seafloor to transport the oil ashore (ignoring economic reasons)? (b) On what parts of the continental slope, if any, do you think pipelines should not be built?
4. Some sedimentary rocks found in the interior of continents are composed of very fine-grained silicate mineral particles with no calcium carbonate and low metal concentrations. Other continental sedimentary rocks are composed of coarse sand grains with little calcium carbonate, and yet others are composed of fine-grained material that is principally calcium carbonate. Describe the characteristics of the locations at which each of these three rock types were originally accumulated as sediments

CRITICAL CONCEPTS REMINDERS

CC1 Density and Layering in Fluids: Fluids, including the oceans, are arranged in layers sorted by their density. Heat sources under the seafloor can heat and reduce the density of any water present, and this heated water rises through the seafloor and up into the water column.

CC2 Isostasy, Eustasy, and Sea Level: Earth's crust floats on the plastic asthenosphere. Sections of crust rise and fall isostatically as temperature changes alter their density. Oceanic crust cools progressively after it is formed and sinks because its density rises. Thus, the seafloor becomes deeper with increasing distance from an oceanic ridge

CC3 Convection and Convection Cells: Fluids that are heated from below, including water within sediments and cracks in the seafloor, rise because their density is decreased. This establishes convection processes. Heated water rising through the seafloor is replaced by colder water that seeps or percolates downward into the seafloor to replace it.

CC4 Particle Size, Sinking, Deposition, and Resuspension: Suspended particles (either in ocean water or in the atmosphere) sink at rates primarily determined by particle size: large particles sink faster than small particles. Once deposited, particles can be resuspended if current (or wind) speeds are high enough. Generally large particles are more difficult to resuspend, although very fine particles may be cohesive and also difficult to resuspend. Sinking and resuspension rates are primary factors in determining the grain size characteristics of beach sands and sediments at any given location.

CC6 Salinity, Temperature, Pressure, and Water Density: Sea water density is controlled by temperature, salinity, and to a lesser extent pressure. Density is higher at lower temperatures, higher salinities, and higher pressures. Heated water discharged by hydrothermal vents has a high enough salinity in some areas of very limited mixing that it is more dense than the overlying seawater and collects in a layer next to the seafloor.

CC7 Radioactivity and Age Dating: Some elements have naturally occurring radioactive (parent) isotopes that decay at precisely known rates to become a different (daughter) isotope. Measuring the concentration ratio of the parent and daughter isotope can be used to calculate the age of the various materials since they were first formed, but only if none of the parent or daughter isotope are gained or lost from the sample during this time period. This condition is usually not met in sediments and radioisotope age dating is difficult so other dating methods including variations in fossil assemblages, and magnetic field properties are used extensively.

CC8 Residence Time: The residence time of seawater in a given segment of the oceans is the average length of time the water spends in the segment. In restricted arms of the sea or where residence time is long, very fine grained particles from glaciers and rivers can become concentrated in the water. Residence time is also a major factor in determining the change in concentration of an element in seawater if its inputs to the oceans change.

CC9 The Global Greenhouse Effect: The oceans play a major part in studies of the greenhouse effect as the oceans store large amounts of carbon dioxide both in solution and as carbonates in sediments, formed at shallow enough depths that the carbonates are not dissolved.

CC14 Phototrophy, Light, and Nutrients: Chemosynthesis and phototrophy)which includes photosynthesis) are the two processes by which simple chemical compounds are made into the organic compounds of living organisms. Many sediments have high concentrations of particles that originate from photosynthetic organisms or species that consume these organisms.

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