4

SEDIMENTS

Objectives

1. Origins and characteristics of sediment particles and deposits;
2. Processes controlling abundances and distributions of sediment deposits in the deep ocean;
3. Sediment transport processes in the ocean and atmosphere;
4. Processes controlling sediment accumulations on continental margins;
5. Methods of deciphering ocean history; and
6. Processes controlling deposits of oil and gas and mineral deposits in the ocean.

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Sediment deposits cover most of the ocean bottom. They are mixtures of particles transported by rivers, glaciers, and winds, and shells and skeletons of marine organisms. A few areas of ocean bottom are devoid of a sediment cover; these are either swept clean by strong bottom currents or too newly formed (midocean ridge crests) to have accumulated sediment deposits. Sediment deposits are usually thickest over the old oceanic crust near passive continental margins, especially in marginal ocean basins.

The ocean bottom is the final repository for insoluble debris from the land. Oceanic sediments record the ocean's history, including climate changes over the past 200 million years. Older sediment deposits have been recycled through subduction or incorporated into the continents.

In this chapter we discuss:

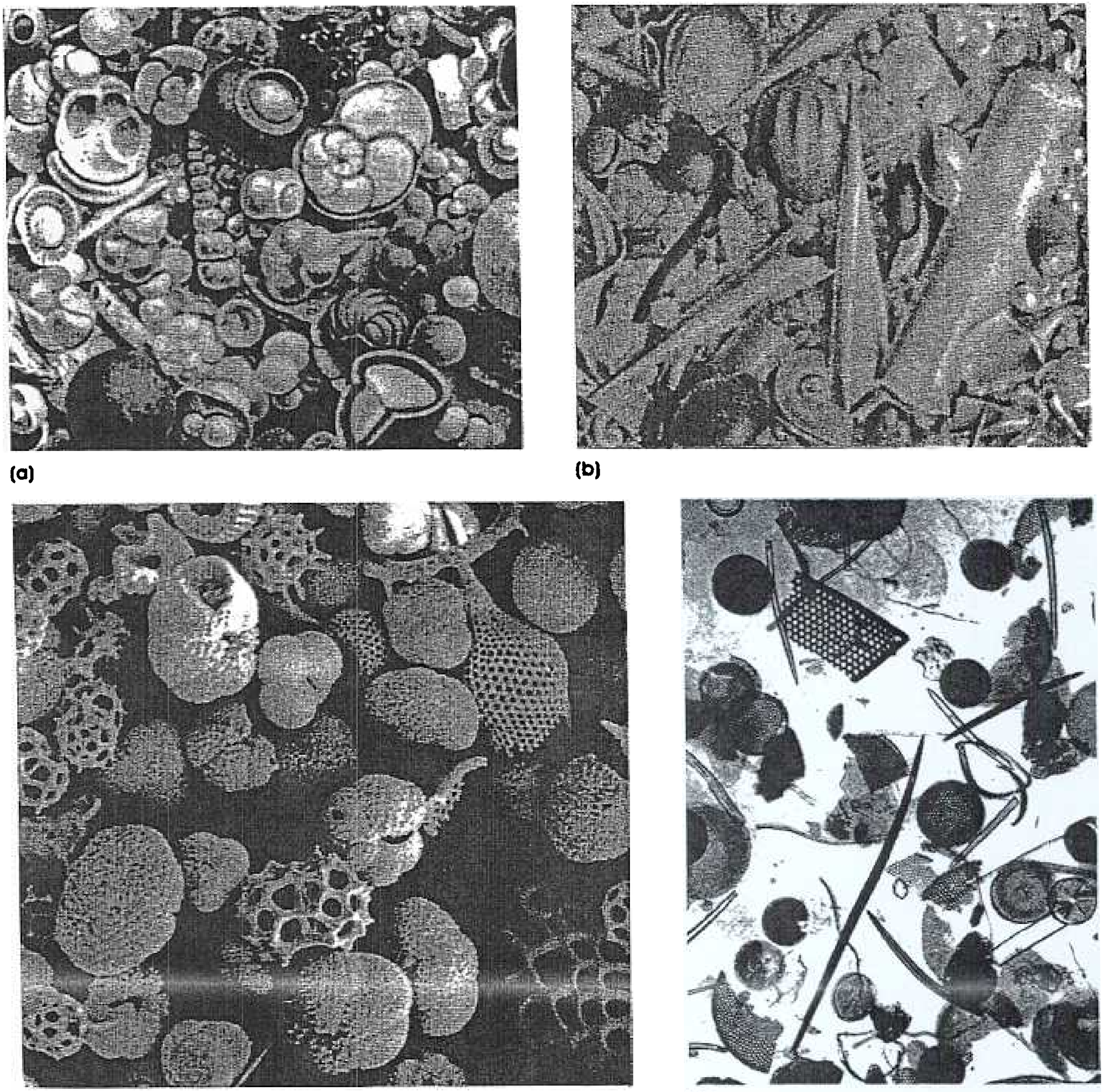
- Origins and characteristics of sediment particles and deposits;
- Processes controlling abundances and distributions of sediment deposits in the deep ocean;
- Sediment transport processes in the ocean and atmosphere;
- Processes controlling sediment accumulations on continental margins;
- Methods of deciphering ocean history; and
- Processes controlling deposits of oil and gas and mineral deposits in the ocean.

ORIGINS AND SIZES OF SEDIMENT PARTICLES

Sediment particles come from three sources:

Biogenous particles come from the bones, shells, and teeth of marine organisms, as illustrated in Fig. 4-1. If a deposit has more than 30% (by volume) biogenous particles, it is called a **biogenous sediment** or sometimes an **ooze**.

Lithogenous particles are rock fragments or minerals released by the decomposition (weathering) of rocks on land and from volcanic eruptions (Fig. 4-2).



(a) (b) (c) (d)

FIGURE 4-1 Constituents of biogenous sediments. (a) Foraminiferal (Globigerina) and (b) pteropod shells in deep-ocean sediments obtained by the *Challenger* Expedition. (c) Skeletons of foraminifera and radiolaria (shells with large openings) from sediment in the western Pacific Ocean. (d) Shells of diatoms.

Hydrogenous particles are formed by chemical reactions occurring in seawater. Manganese nodules, an example of hydrogenous sediments, are illustrated in Fig. 4-3.

Sediment particles are also classified according to grain size, as shown in Fig. 4-4. Particle size is important because it determines the

FIGURE 4-2
Volcanic eruptions inject large volumes of ash into the atmosphere. Winds transport ash for long distances, so much ash is deposited in the deep ocean. This ash cloud was caused by an eruption of La Soufrière on April 17, 1979, on the island of St. Vincent in the West Indies. (Courtesy NASA.)

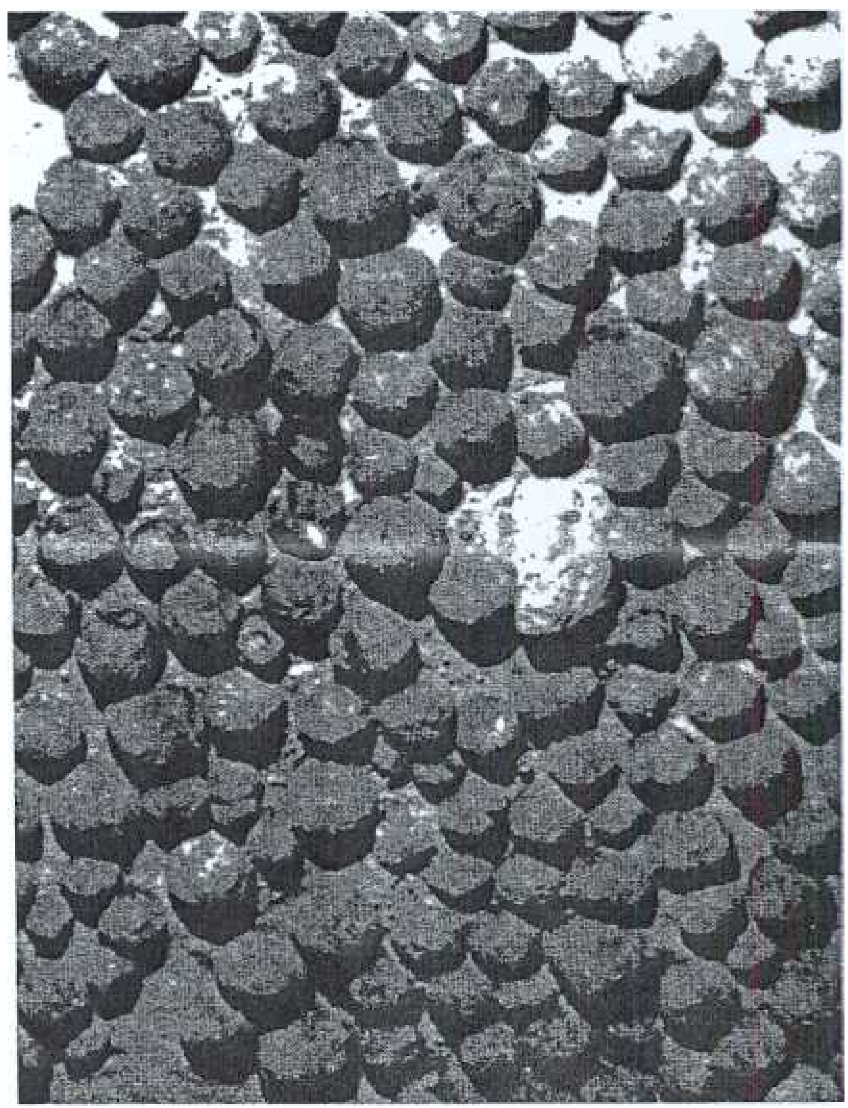


FIGURE 4-3
Manganese nodules on the floor of the South Pacific. These nodules are 8 to 10 centimeters across. (Courtesy National Science Foundation.)

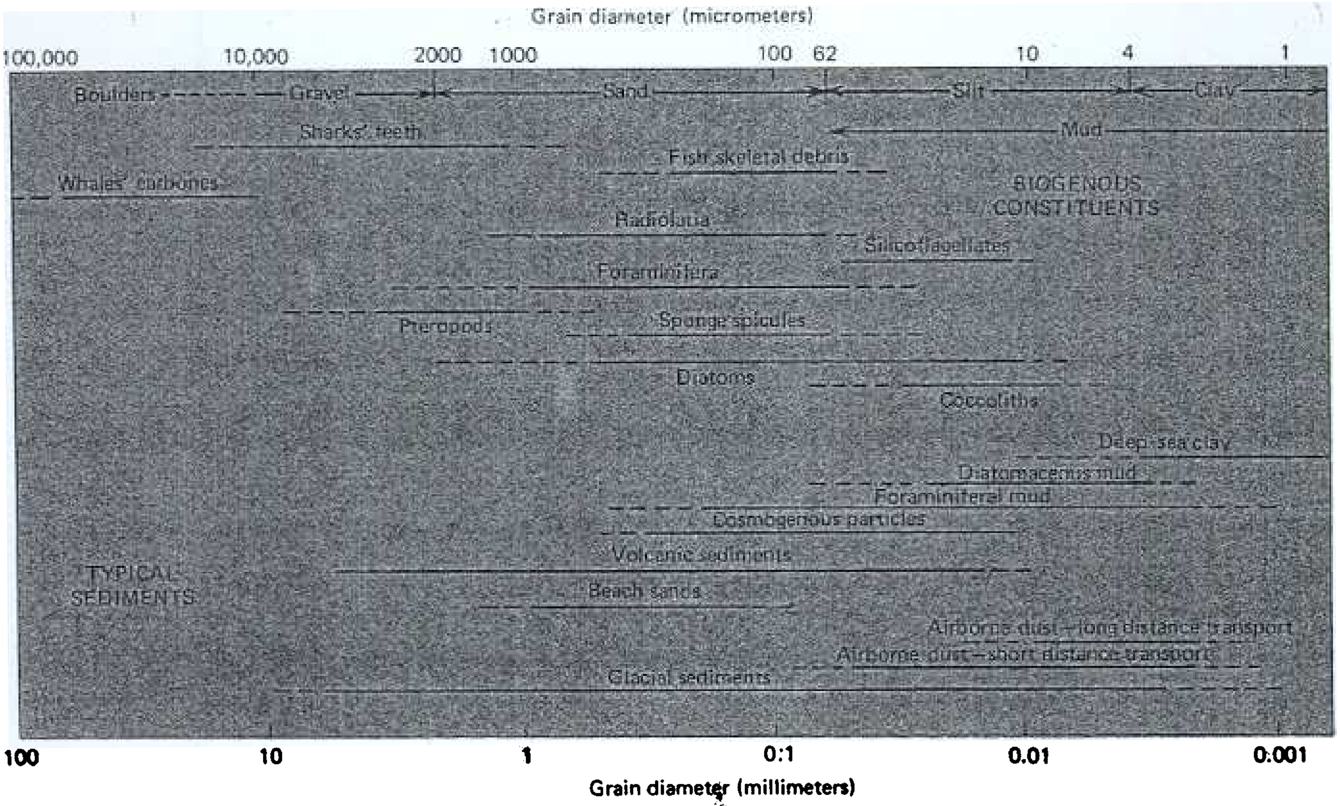


FIGURE 4-4
Grain-size distributions in common marine sediments and in various sediment sources.

primary mode of particle transport and the distances particles travel through the air or ocean before settling out to be deposited on the ocean bottom.

Various sediment sources produce particles of different sizes, as indicated in Fig. 4-4. For instance, glaciers transport particles that range in size from boulders to fine-grained rock flour, which is produced as the glaciers move across the land. Sediments with this wide a range of particle sizes are called **poorly sorted**. On the other hand, ocean beaches usually consist of **well-sorted** sands because wave action has removed the fine particles. Deep-ocean deposits usually consist of poorly sorted mixtures of fine-grained particles. These we call **deep-sea muds**.

BIOGENOUS SEDIMENTS

Sediment deposits covering more than half the deep-ocean bottom are dominated by the remains of tiny floating plants and animals either calcareous or siliceous, as indicated in Fig. 4-5. Distribution of these biogenous sediments on the deep-ocean floor is governed by three processes: **biological production** in overlying surface waters, **dilution** by other kinds of sediment particles, and **destruction** of particles, primarily by chemical processes.

Highly productive waters are often underlain by biogenous sediments. Under the highly productive equatorial waters of the Pacific and Indian oceans, deposits rich in radiolarians (a one-celled animal, discussed in Chapter 13) accumulate, up to 1 kilometer thick in some areas. Diatomaceous sediments occur in the high-latitude areas of the North Pacific and near Antarctica (see Fig. 4-5) where there are short, intense bursts of diatom growth. Diatoms are one-celled floating plants (also discussed in Chapter 13).

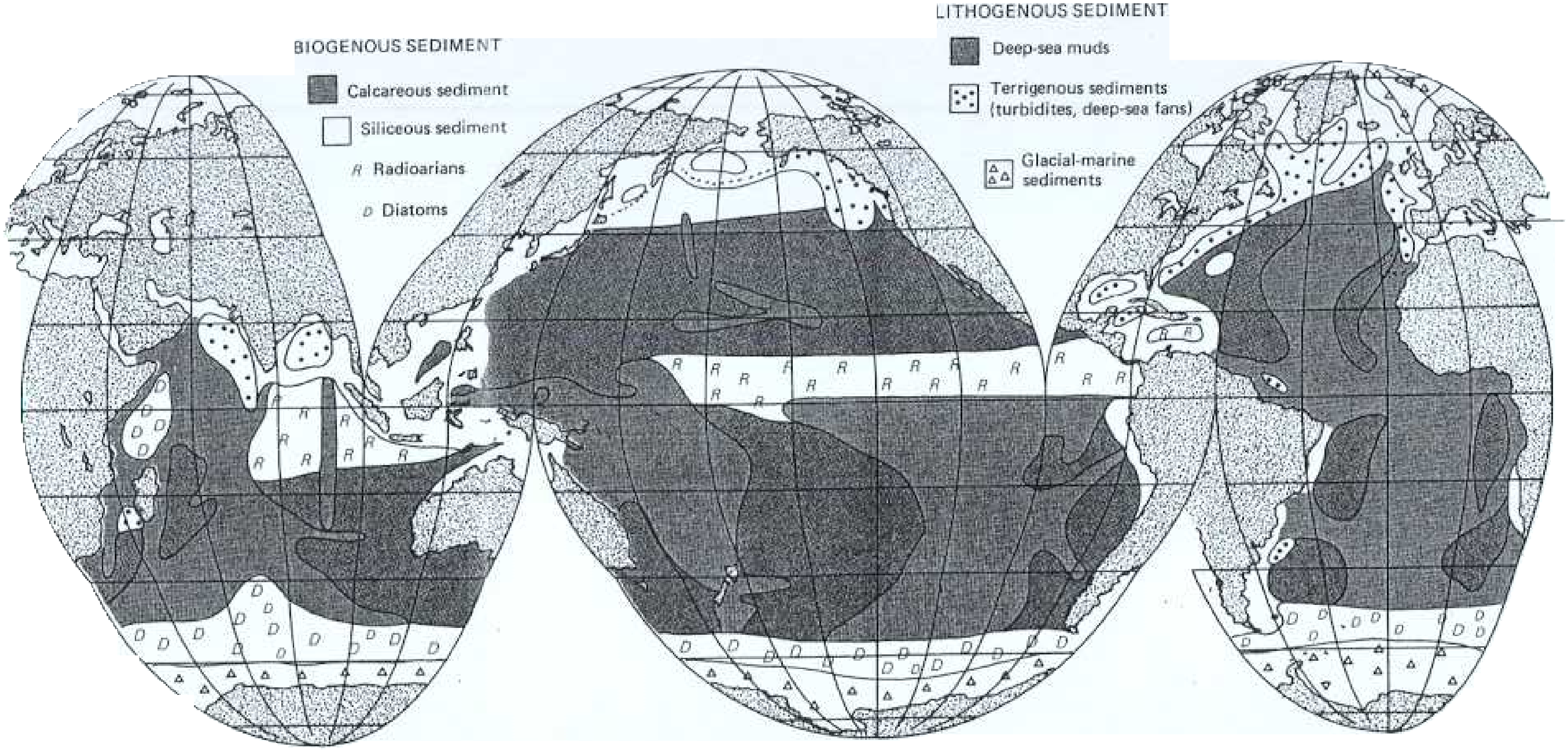


FIGURE 4-5 Distributions of deep-sea sediment deposits. Continental shelf and slope sediments are not shown.

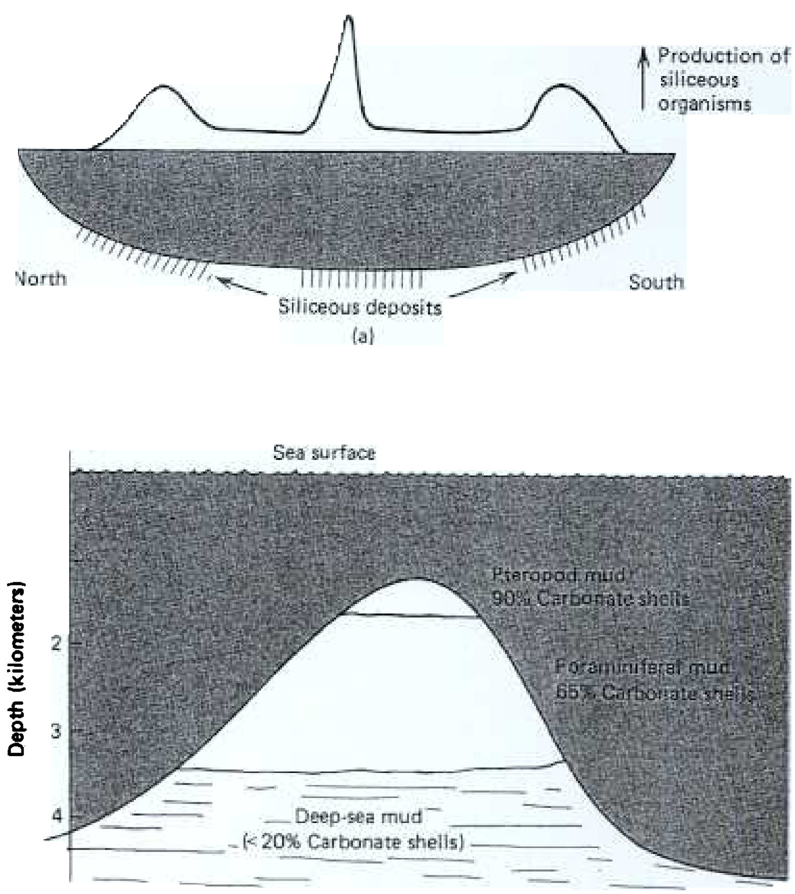
Biogenous particles are mixed with lithogenous particles during deposition. Note in Fig. 4-5 that diatomaceous sediments in the North Atlantic are masked by large amounts of sediments coming from the land. Large amounts of rock debris eroded by Antarctic glaciers and released by melting ice dominate sediment deposits there.

DESTRUCTION OF BIOGENOUS PARTICLES

Shells, bones, and teeth dissolve rather quickly in seawater. Thus deep-sea deposits of these materials commonly occur primarily in areas where plants and animals grow in abundance [Fig. 4-6(a)]. Because of their rapid dissolution, only fragments of dense fish bones and resistant fish scales or sturdy siliceous shells survive in most marine deposits. Some of the more resistant particles, such as fish teeth or whale ear bones, persist for a long time on the ocean floor.

Carbonate particles dissolve fairly slowly in shallow- and intermediate-depth waters. Calcium carbonate minerals become more soluble at the relatively low temperatures and high pressures of deep-ocean waters. The depth at which carbonate particles will dissolve is called the **carbonate saturation level**; it occurs at around 4000 meters, as shown in Fig. 4-7, where the rate of weight loss for carbonate shells increases markedly. Below the carbonate saturation level, carbonate particles dissolve before they form calcareous sediments, except in areas of extremely high production of shells.

FIGURE 4-6
(a) Distributions of siliceous deposits are related to ocean areas of high productivity, such as the equatorial upwelling region and high latitudes. Compare with Fig. 4-5. (b) Depth zonation of calcareous sediments on a seamount in the deep ocean.



(b)

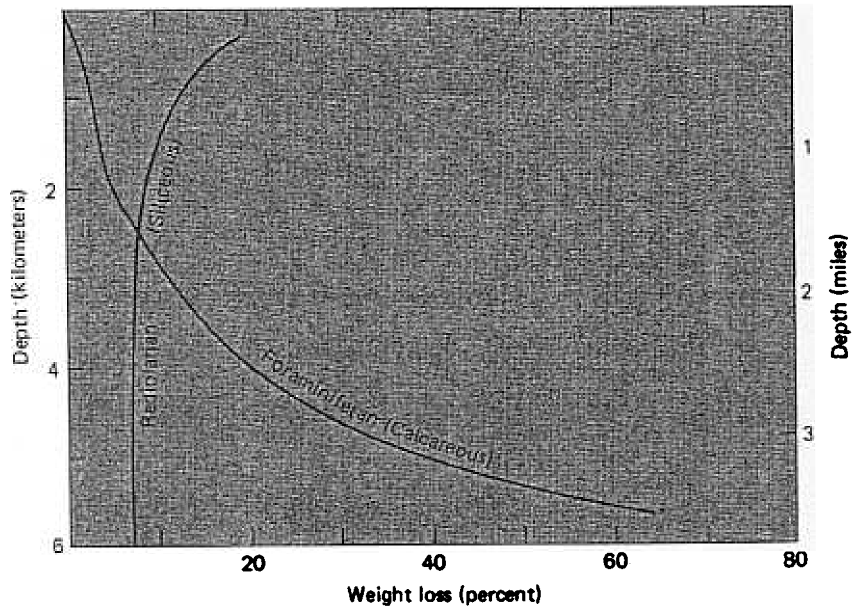


FIGURE 4-7
Loss of weight of particles due to dissolution at various depths in the central Pacific Ocean after 4 months of submergence.

Some organisms, such as pteropods (a small floating snail, discussed in Chapter 13), form their carbonate shells from an especially soluble carbonate mineral; hence, pteropod shells dissolve at much shallower depths than shells formed of less soluble carbonate minerals. Pteropod shells form deposits only on shallow volcanic peaks in the Atlantic [see Fig. 4-6(b)].

LITHOGENOUS SEDIMENTS

Lithogenous particles come from the breakdown of silicate rocks on land. Most rocks originally formed at high temperatures and pressures in the absence of free oxygen and with little liquid water. These rocks break down at the earth's surface due to chemical and physical processes known as **weathering**. Soluble constituents are released and carried to the ocean, dissolved in river water. The remaining rock is broken up by physical processes, such as freezing, into sand- and clay-sized grains, which are also carried suspended in running water or by wind.

Some minerals—for example, quartz—resist chemical alteration. Usually entering the ocean almost unaltered, they are common in sands and silts. The slightly altered remains of micalike minerals form clay minerals, common in fine-grained deep-sea clays.

Rivers transport most of the lithogenous sediment to the ocean—about 13 billion tons each year, most of it coming from Asia, as shown in Fig. 4-8. The four largest rivers are in Asia; together they supply about 25% of the total sediment discharge from continents (see Fig. 4-8). An unknown but large amount of sediment is carried by winds into the deep-ocean basins.

Except for the Ganges-Brahmaputra rivers (which discharge into the northern Indian Ocean) and the Amazon and Congo rivers (which drain into the equatorial Atlantic) most major sediment-producing rivers drain into marginal seas. Consequently, about two-thirds of the river-borne sediment discharged does not reach the deep-ocean floor.

Much lithogenous sediment comes from semiarid regions, and especially from areas of active mountain building. Normally rainfall in such areas is inadequate to support an erosion-resisting cover of

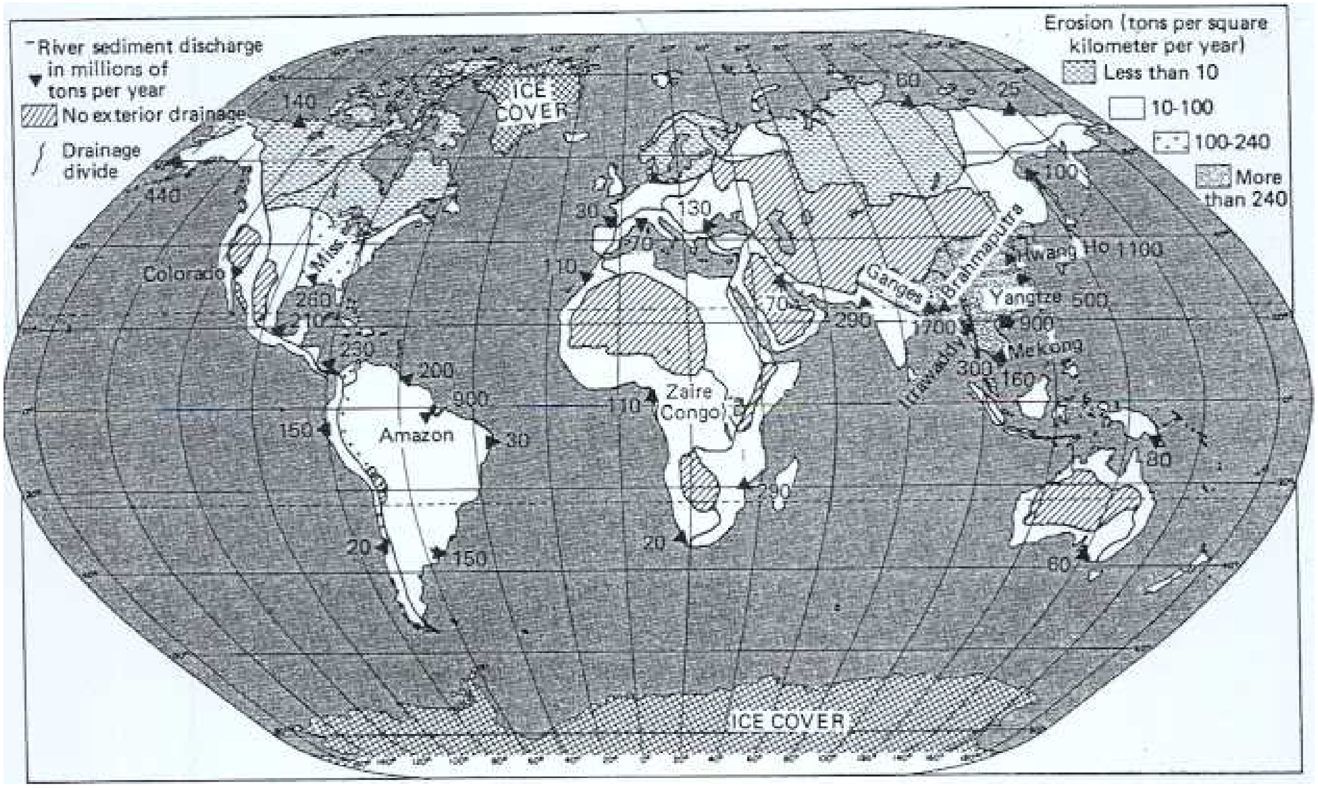


FIGURE 4-8
Major sources of river-borne lithogenous sediment.

vegetation but is sufficient to erode soils. In Asia and India active mountain building causes rapid erosion and high sediment discharges. Long-continued intensive agriculture has further increased sediment yield.

HYDROGENOUS SEDIMENTS

Hydrogenous constituents are formed by chemical and biological processes on the ocean bottom. Potato-sized nodules of iron and manganese (see Fig. 4-5) are conspicuous examples of hydrogenous sediments. Iron and manganese are brought to the ocean by rivers and by hydrothermal vent discharges. Both elements form tiny particles in seawater which are swept along by currents until they strike a surface, where they stick, forming a stain or crust.

In the centers of ocean basins, far from continents, particles accumulate very slowly. There nodules form by the accretion of hydrogenous particles around rocks, whale ear bones, or shark teeth. Over millions of years, iron–manganese coatings also collect other types of sediment particles as impurities, eventually forming a nodule with roughly concentric rings (see Fig. 4-9). Such nodules contain relatively high copper, cobalt, and nickel concentrations, which make them attractive as potential sources for the extraction of these metals. Similar accumulations coat the rocks on seamounts and volcanoes in the central Pacific.

Because of their slow growth, manganese nodules cannot form or are buried where sediments accumulate rapidly. Consequently, nodules are rare in the Atlantic, which receives large amounts of sediment discharged by many rivers. But in the central Pacific, where sediments accumulate very slowly, nodules cover an estimated 20 to 50% of the ocean bottom. Tunneling and churning by sediment-feeding organisms

FIGURE 4-9
Growth rings of a manganese nodule are seen in this cross section. The rings are usually formed by tubes and other biological structures built by organisms living on the nodule surface during its formation.

may also help to keep nodules from being buried, so they continue to grow at the sediment surface.

SEDIMENT TRANSPORT

Particle transport is controlled primarily by physical processes. Particle size and current speed are the most important properties determining sediment transport (Fig. 4-10). Vertical movements of particles as they sink are the easiest to visualize. Settling speed is controlled primarily by particle diameter; large particles sink more rapidly than small ones. Settling times for lithogenous particles are given in tabular form.

PARTICLE (diameter in micrometers)	SETTLING VELOCITY (centimeters per second)	TIME TO SETTLE 4 KILOMETERS
Sand (100 μm)	2.5	1.8 days
Silt (10 μm)	0.025	185 days
Clay (1 μm)	0.00025	50 years

Large sand grains settle out rapidly near the place where they enter the ocean. Finer grained silts can be transported greater distances during the 185 days it takes them to settle 4 kilometers. Clay particles can be transported throughout the ocean during the 50 years they take to settle through the ocean as single particles. Settling times can also be increased by upward-directed water movements resulting from turbulence. Silts and clays are also transported hundreds or thousands of kilometers by winds.

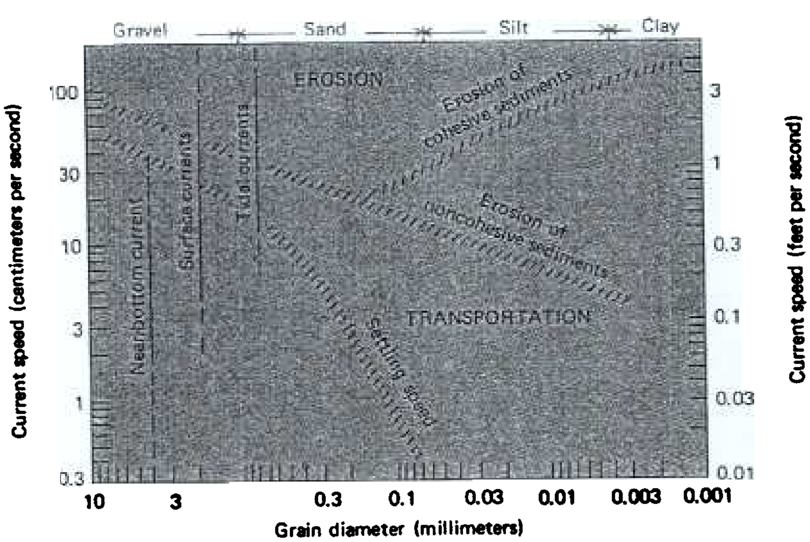
Let us now consider how currents erode and transport sediment in the ocean. As Fig. 4-10 shows, current speeds necessary to erode and suspend noncohesive deposits (such as sand) decrease for smaller particles; even weak currents can erode and transport fine-grained noncohesive sediments. But very fine-grained particles are also usually cohesive, so forces necessary to erode them increase with the smaller-grained materials. Thus the currents required to move cohesive silts and clays are as strong as those needed to move sands and gravels. Even on the deep-ocean bottom, currents are often strong enough to erode and move sediments. This is especially common under the strong currents along the western ocean boundaries.

Organisms also cause sediment deposition by filtering water. Mineral particles removed along with their food are compacted into fecal pellets and then excreted; they sink rapidly (within a few days) to the bottom. Oysters, mussels, clams, and many other attached and floating animals effectively remove particles from suspension and bind them into fecal pellets that accumulate nearby. Plants growing on tidal flats also trap sediment by providing quite sheltered areas where these small particles can settle out. Plants then prevent erosion of sediment deposits when tidal currents or waves scour the area.

Most sediment particles are trapped and deposited near river mouths. Estuaries that have not been filled by sediment act as effective sediment traps; for example, rivers of the U.S. Atlantic Coast transport about 20 million metric tons of sediment each year to their estuaries, but virtually none reaches the continental shelf. Thus the continental shelf of eastern North America is covered primarily by deposits that formed when sea level was lower. There are only small, isolated areas of sediment deposits forming today.

When a river's sediment load is large, it can fill its estuary; then the sediment load can be carried out into the ocean. The Mississippi River is an example. Its large sediment load has not only filled the estuary but also created the Mississippi delta, a massive sediment deposit that extends far out into the Gulf of Mexico (discussed in Chapter 11).

FIGURE 4-10
Current speeds required to erode sediment deposits and to transport grains as compared to settling rates for individual grains. Note that the scales for grain size and current speed increase by factors of 10.



ATMOSPHERIC TRANSPORT

Perhaps as much as 100 million metric tons of lithogenous particles per year are transported by winds to the ocean, mainly from deserts and high mountains having sparse vegetation and strong winds (see Fig. 4-13). Wind transport of sediment to the deep ocean seems to be especially important in middle latitudes, where there are few rivers, owing to the arid climate. It is less important in the Southern Hemisphere, where there is less land.

Particles less than 20 micrometers in diameter may be carried great distances by winds. Volcanic fragments smaller than about 10 micrometers may be carried around the world if an eruption is powerful enough to inject fine ash into the stratosphere.

Particles eroded by winds remain in the lower atmosphere, forming a "plume" of such material. This situation is particularly noticeable off Africa, where Sahara sand is often blown hundreds of kilometers to sea (Fig. 4-11). Rust-covered grains from the Sahara are common in North Atlantic sediment deposits.

Immense volumes of volcanic ash from large volcanic eruptions fall into the ocean each year. Very large eruptions are infrequent. The Indonesian volcano Tambora in 1815 released nearly 80 cubic kilometers of ash and left an easily recognizable halo of volcanic ash in sediment deposits around the volcano. On August 26, 1883, Krakatoa, an island between Java and Sumatra in Indonesia, erupted. About 16 cubic kilometers of ash were discharged to the atmosphere, much of it going into the stratosphere. An area of nearly 4 million square kilometers had ash deposits, and large amounts were deposited on the sea floor. The May 18, 1980, eruption of Mt. St. Helens in Washington injected almost 1 cubic kilometer of ash into the atmosphere.

FIGURE 4-11
Dust from the Sahara Desert was blown over the North Atlantic. On August 5, 1974, a dust cloud reached all the way to Florida.

Because of the enormous releases of energy during these volcanic eruptions, large amounts of ash are injected into the upper atmosphere (stratosphere) to be transported around the world. Ash from Krakatoa was observed for several years over Europe, causing spectacular sunsets.

DISTRIBUTION OF DEEP-SEA SEDIMENTS

Much of the deep-ocean floor is covered by **pelagic sediments** that accumulate slowly, particle by particle, at rates between 1 and 10 millimeters per thousand years. Deposits formed in this way blanket the original ocean bottom topography, faintly preserving its outlines. A comparison has often been made with newly fallen snow on land.

Deep-ocean sediments accumulate slowly (see Table 4-1). Thus the particles not only spend years suspended in ocean water but also remain exposed on the ocean bottom for centuries before they are finally buried and sealed off from contact with near-bottom waters. As a result, there is usually ample time for particles in deep-ocean sediments to react with seawater. Abundant dissolved oxygen in deep-ocean waters causes iron to be converted to the ferric state (iron rust), giving red clays their typical color.

Thinnest sediment deposits generally occur on the young crust of midoceanic ridges or rises. Large areas may also be swept clear of sediment by strong currents; sediments accumulate there in protected "pockets." Sediment thickness increases away from the young ridges and is generally greatest over the oldest crust near the continents. Sediments form abyssal plains along the ocean basin margins.

TABLE 4-1
Typical Sediment Accumulation Rates

AREA	AVERAGE (RANGE) ACCUMULATION RATE (centimeters per thousand years)
Continental margin	
Continental shelf	30 (15–40)
Continental slope	20
Fjord (Saanich Inlet, British Columbia)	400
Fraser River delta (British Columbia)	700,000
Upper Gulf of Thailand	400–1100
Marginal ocean basins	
Black Sea	30
Gulf of California	100
Gulf of Mexico	10
Clyde Sea	500
Deep-ocean sediments	
Coccolith mud	1 (0.2–3)
Deep-sea muds	0.1 (0.05–0.8)

CLAYS

Deep-sea muds cover the deeper parts of the ocean basin. They are dominated by clay particles, which are less than 4 micrometers in diameter (see Fig. 4-4). Over most of the central ocean basins, far from land, these particles are carried by winds.

Clay minerals have a distinctive layered structure, much like the common micas one sees in rocks on land. Different minerals are formed under different climatic conditions, so their presence in deep-sea deposits provides clues to the sources of the particles. For example, one group of clays called **kaolinite** forms during weathering of rocks under tropical conditions. This is reflected in their distribution in modern marine sediments, where they are restricted to tropical waters (Fig. 4-12).

Other clays are formed when rocks are weathered in the mid-latitudes. These minerals, called **illite**, are carried by winds from the continents and deposited in the Pacific and North Atlantic, where they are especially abundant (Fig. 4-13).

TURBIDITY CURRENTS

A turbidity current is a dense mixture of water and sediment that flows along the bottom, transporting sediment [see Fig. 4-14(a)] from continental margins onto the deep-ocean floor. The best way to visualize a turbidity current is as a huge watery landslide.

Earthquakes can cause them. In November 1929, a turbidity current triggered by the Grand Banks earthquake hurtled down the continental slope off Newfoundland, breaking telegraph cables along its way. The flow moved like a gigantic submarine avalanche at speeds of up to 100 kilometers per hour, finally depositing sediment over a large area of ocean bottom. A sudden large discharge of river-borne sediment (during a flood) can trigger a turbidity current.

When the current first forms, it moves at high speeds and carries sand-sized particles eroding deposits as it passes. Later the flow moves progressively more slowly, transporting finer particles that are deposited on top of the heavy ones. The result is a rapidly formed sediment deposit showing distinct gradation in particle size, changing from coarsest grained at the bottom to the finest at the top, called **graded bedding** [see Fig. 4-14(b)]. Turbidity currents carry shells of shallow-water organisms and plant fragments from shallow waters onto the deep-ocean floor. Plant fragments—some still green—have been dredged

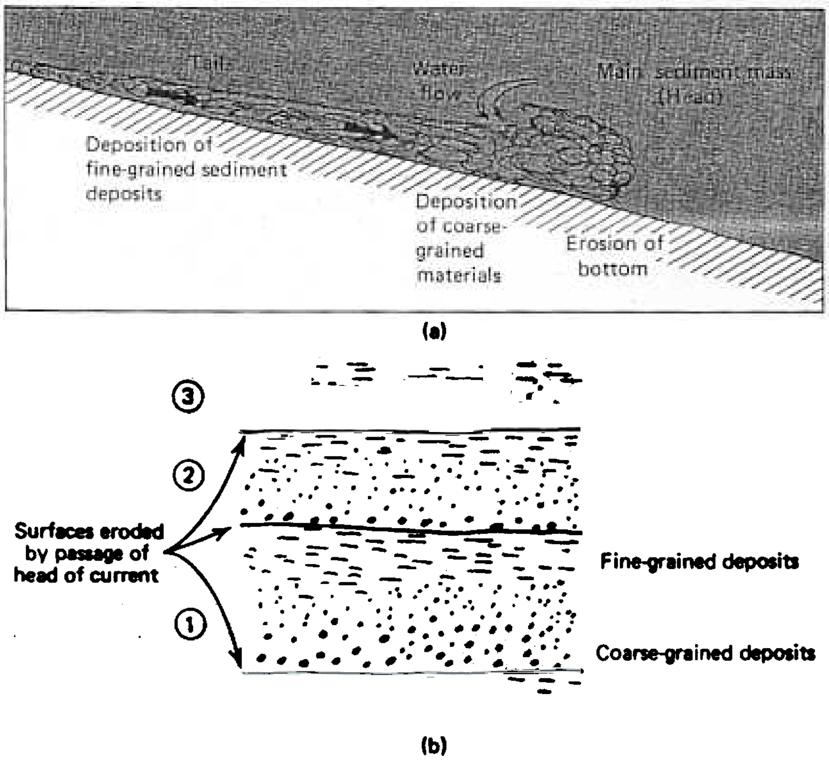


FIGURE 4-14
(a) Schematic representation of a turbidity current moving down a slope. Note that the current first erodes as it passes. As current speeds decrease when the main sediment mass passes, coarse-grained sediments are deposited first; fine-grained sediments are deposited later. (b) Deposits formed by turbidity currents.

from the ocean bottom following cable breaks.

Turbidity currents are most active on narrow continental shelves, and least active on wide continental shelves. In the Atlantic, turbidity current deposits have buried most of the older topography, so currents flow from the edge of the continents almost to the midocean ridge. Similarly, sediment brought into the northern Indian Ocean can flow unimpeded over great distances of deep-ocean bottom. But in the Pacific, trenches and ridges along the ocean basin margins prevent movement of turbidity currents onto the deep-ocean floor.

Besides carrying sediment and burying ocean bottom topography, turbidity currents also cut submarine canyons that indent most continental margins. Whether turbidity currents are the sole agent responsible is still not known. There is good reason to suppose, however, that in many areas they serve to remove the large volumes of sediment that would otherwise fill these canyons completely. Evidence for this comes from frequent breakage of submarine cables off the mouths of large rivers, such as the Zaire (Congo) River.

DISTRIBUTION OF CONTINENTAL MARGIN DEPOSITS

Transport processes control marine sediment distribution. In general, river-transported sediment is restricted to continental margins. Most sediment remains on the shelf or rise except where turbidity currents carry material out into the deep ocean.

Most ocean margins are covered by thick deposits of lithogenous sediment. Transport of this material from continents to the deep-ocean bottom is inhibited by the topography of continental margins. The lack of modern sediment brought to the continental shelf by rivers is a result of sea level changes following the last retreat of the glaciers (sea level changes are discussed in Chapter 11).

Deposits of sand and silt that were formed under conditions no longer existing in an area are known as **relict sediments**. They have distinctive features, such as shells of organisms that no longer grow there. Oyster shells far out on the U.S. continental shelf, for example, indicate deposition at a time of much lower sea level. Mastodon teeth and remains of extinct land animals indicate that the area was once dry and that its surficial deposits formed at that time. Other characteristics of relict sediments include iron stains or coatings on grains that could not have been formed under marine conditions.

Where rivers bring sediment to the coastal ocean, relict deposits are buried. Alternatively, they may be reworked by wave action when sea level is rising, so that the characteristic features of a relict deposit are destroyed or masked. About 70% of the world's continental shelves are now covered by relict sediments (Fig. 4-15).

Sand deposited near the mouth of a river is moved by longshore currents. The association of sand beaches and river mouths is especially obvious on the U.S. Pacific Coast. There the largest beaches and dunes in Washington and Oregon are associated with the Columbia River mouth.

Sands are common in areas of strong wave action because finer-grained materials are removed. Silts and clays are carried farther seaward and deposited at depths where wave action is too weak to stir up sediment or erode the bottom. On many continental shelves this process occurs at depths of 50 to 150 meters.

Sands and silts deposited on continental shelves accumulate so rapidly that particles have too little time to react chemically with the

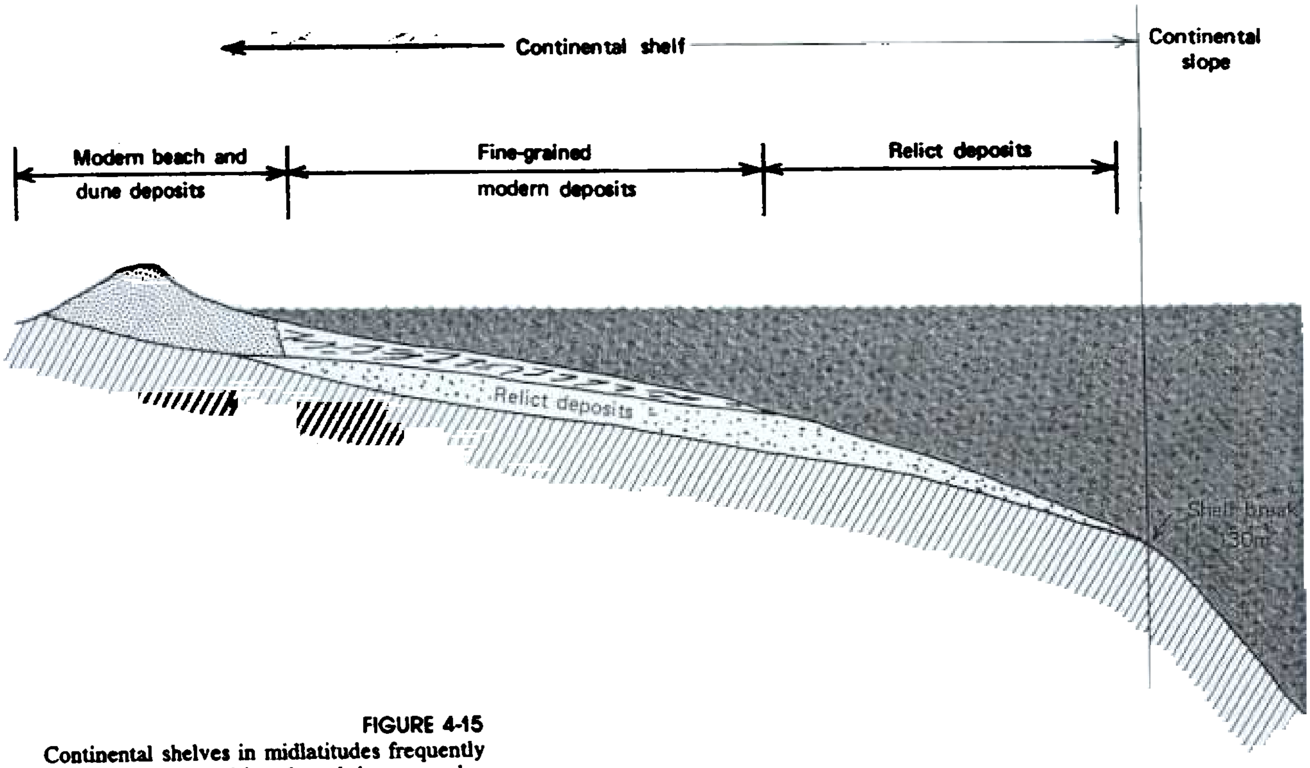


FIGURE 4-15
Continental shelves in midlatitudes frequently have an active sand beach and dune complex along the shore that merges with a deposit of finer-grained modern sediments at mid-depths on the shelf. At the outer edge of the shelf are much older deposits.

seawater or with near-bottom waters as they lie on the ocean bottom before being buried. Coastal ocean sediments therefore retain many characteristics acquired during weathering. These rapidly accumulating deposits tend to be rather dark colored—gray, greenish, or sometimes brownish—and they contain abundant organic matter.

In polar regions, glacial marine sediments are common. These deposits contain particles of all sizes—from boulders to silt (see Fig. 4-4) deposited by melting icebergs from continental glaciers. Sea ice normally contains sediment only if it has gone aground and incorporated material from the shallow bottom. This happens in the Arctic Sea. When such sea ice melts, its sediment load is deposited on the bottom, where it mixes with remains of marine organisms.

Near active volcanoes, the ocean receives substantial amounts of volcanic rock fragments, usually the relatively large particles (greater than 50 micrometers in diameter) formed during eruptions. Sediments deposited in the Indonesian area, around the Aleutians off Alaska, and near western Pacific island arcs contain abundant volcanic debris. Volcanic sediments are not restricted to the Pacific Ocean; ash occurs in all ocean basins, transported by winds from major volcanic eruptions.

Sediments that accumulate fairly rapidly near continents (see Table 4-2), transported either by running water, turbidity currents, or ice, cover about 25% of the ocean bottom. Accumulations are thickest in marginal ocean basins; although accounting for only about 2% of the ocean area, they contain about one-sixth of all oceanic sediment.

Some general relationships between sediment-transport processes and sediment distributions are summarized diagrammatically in Fig. 4-16. Effects of high productivity can be seen on the deep-ocean bottom and on the continental shelf. On the deep-ocean bottom (Fig. 4-16) bands of diatomaceous muds in high latitudes and radiolarian muds in equatorial regions directly reflect the high biological productivity of surface waters in these regions. On the continental shelf the abundance of recent carbonate sediment in tropical waters is again a result of the locally high carbonate productivity. Figure 4-16 shows that turbidity currents leave thick sediment deposits on and near continental margins.