CHAPTER 1 INTRODUCTION

> 'He had long ago decided, since he was a serious scholar, that the caves of ocean bear no gems, but only soggy glub and great gobs of mucky gump.'
>
> James Thurber

Most substances that enter the oceans ultimately end up in the sediments. On the way, they participate in a variety of complex biological and chemical cycles and interactions which involve some substances more than others. Interactions continue after deposition: sediments do not lie passively on the sea-bed until they are buried. Deep-sea animals disturb the sediments as they forage for food, and some sediment may experience erosion and resuspension by bottom currents before being redeposited and finally buried. Chemical reactions may occur between the mineral grains and the overlying seawater, and these reactions can continue after burial, when seawater becomes trapped among the grains.

Nowadays, deep-sea sediments are a focus of much research effort, because of the growing need to quantify the various fluxes contributing to the global carbon cycle. A principal aim is to find out just what happens to the anthropogenic carbon dioxide and other gases contributing to the **greenhouse effect**, which may already have begun to lead to global atmospheric warming. If the present trend continues, major climatic changes and rising sea-levels will result. To predict the extent of such changes successfully, it is essential to know more about rates of increase of greenhouse gases in the atmosphere. This in turn requires improved knowledge of, for example, seasonal, inter-annual and regional fluctuations in ocean surface productivity, the removal of organic matter to the sea-bed, exchanges across the air–sea interface, and vertical and horizontal water movements. International and national programmes initiated to investigate these and related problems include the Joint Global Ocean Flux Study (JGOFS), the Biogeochemical Ocean Flux Study (BOFS), and the somewhat broader **World Ocean Circulation Experiment (WOCE)**.

Nor should the seawater itself be forgotten. The dissolved constituents of seawater include metals such as copper, lead, zinc, tin, manganese, cadmium, mercury, nickel and silver. Geochemical cycles of these and other heavy metals are being grossly perturbed by human activity. Inputs have increased substantially since the Industrial Revolution, although only in the case of lead has the average concentration in open ocean waters actually increased, as a result of atmospheric fall-out. In addition, there are many new substances in the marine environment that were not there even a century ago. These include pesticides and other organic chemicals, as well as transuranic and other 'man-made' nuclides from nuclear weapons testing and low-level waste discharges. Global flux studies will contribute greatly to our understanding of how the many constituents of seawater—both dissolved and particulate, both natural and artificial—move through the various marine chemical cycles. But we begin by looking at the sediments.

1.1 THE DISTRIBUTION OF DEEP-SEA SEDIMENTS

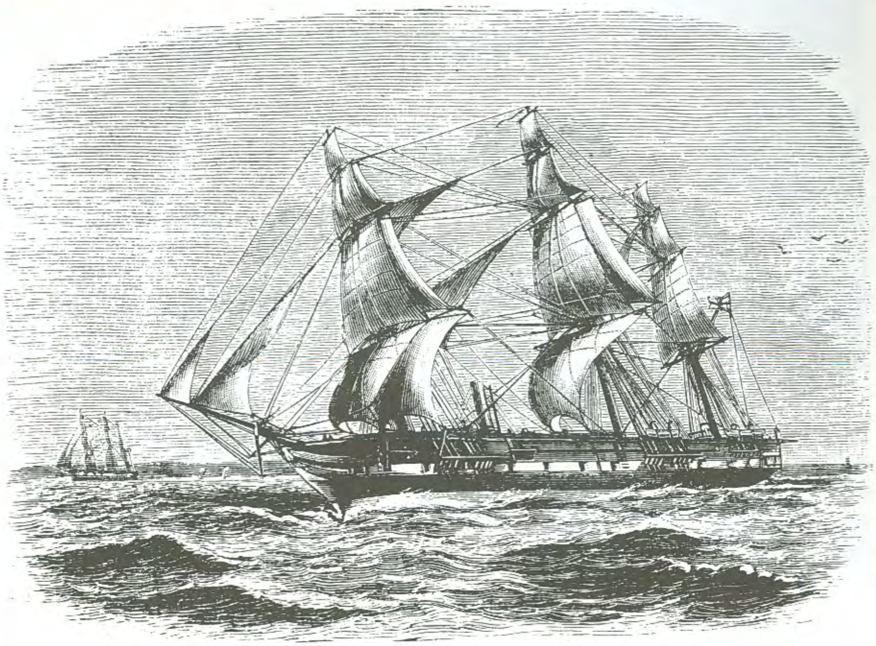
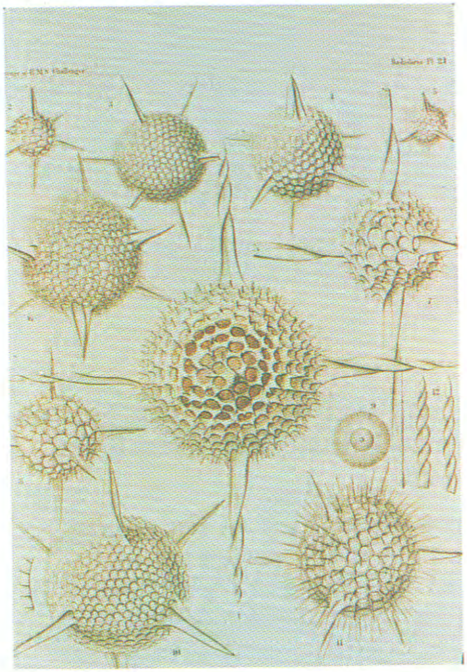


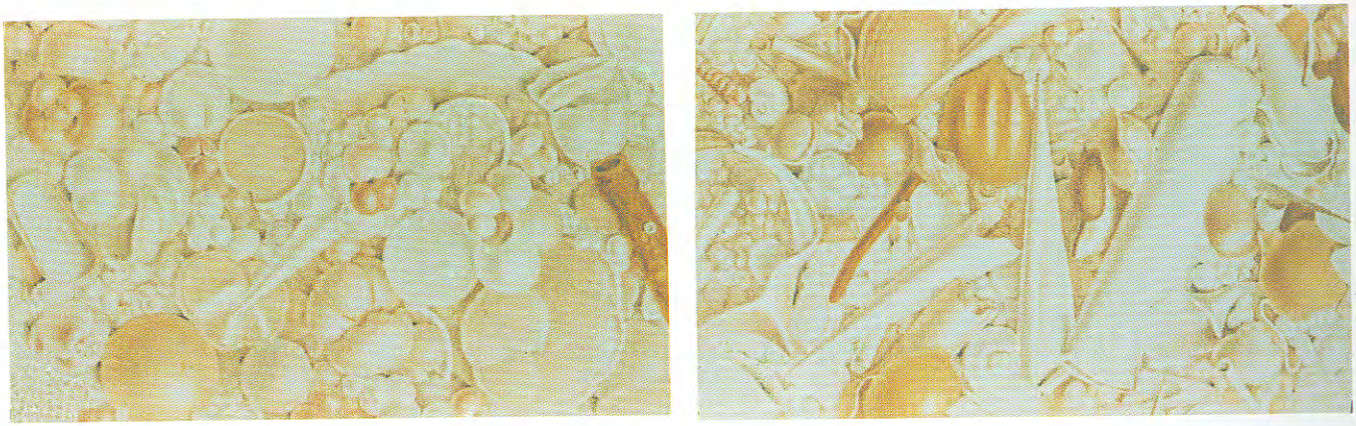
Figure 1.1 HMS *Challenger*, 1872. She was a steam-assisted wooden corvette of 2306 tonnes.

When HMS *Challenger* (Figure 1.1) returned to England on 24 May 1876 laden with specimens, records and measurements after an epic three years and nine months voyage of oceanic exploration, the era of systematic oceanography had begun. The member of the scientific party who did most to ensure world recognition of the *Challenger*'s scientific achievements was John Murray, a Canadian-born Scot, who owed his place on the ship to chance, when a member of the original team was obliged to drop out at short notice. Murray's account of the samples collected from the floor of the oceans (Figure 1.2) provided the starting point for all subsequent investigations into deep-sea sediments.

Figure 1.2 Drawings of planktonic remains in sediments collected from the North Atlantic during *Challenger's* voyage.



(a) Drawings of Radiolaria of the genus *Hexastylus*.



(b) Washed foraminiferal ooze, mainly *Globigerina* spp. (hence the term *Globigerina* ooze). Other shells include a pteropod (centre, see (c)) and an ostracod (crustacean bivalve, extreme left centre). The brown tube (right centre) was probably secreted by an agglutinating (arenaceous) foraminiferan species. Width of field about 3.5mm. Sample from 1900 fathoms (*c.* 3500m).

(c) Washed pteropod ooze. The large cone-shaped shell and some of the larger coiled shells are from pteropods (planktonic gastropods). Foraminiferan remains (see (b)) make up the rest. Width of field 18mm. Sample from 450 fathoms (*c.* 800m).

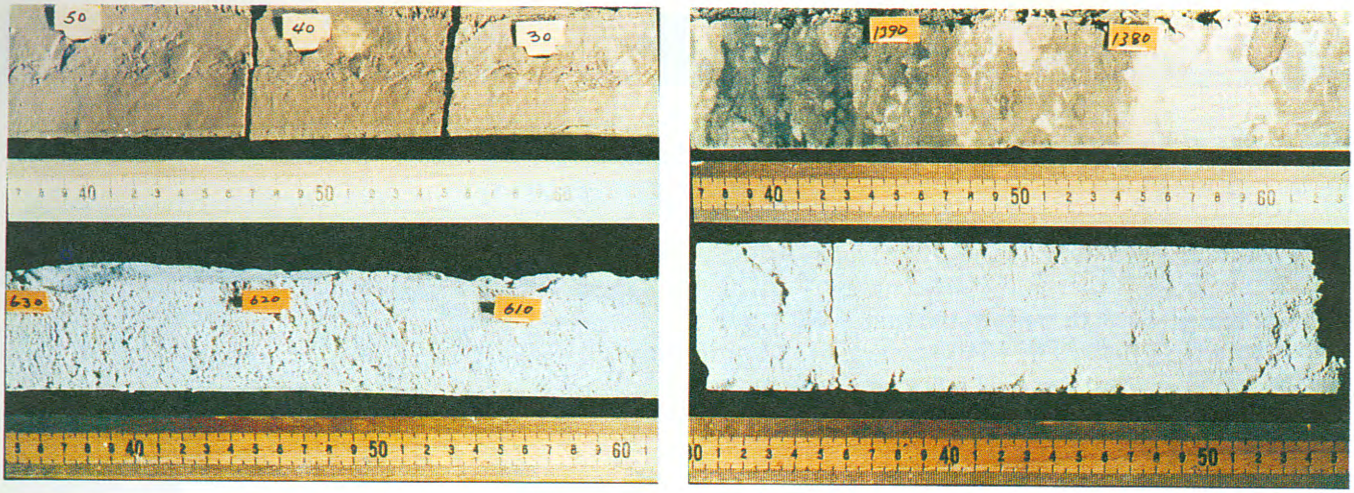


Figure 1.3 Some deep-sea sediment cores,

(a) *Top:* Chocolate-coloured terrigenous 'red clay', typically structureless and with few organic remains other than occasional fish teeth.

Bottom: Diatom ooze from the Antarctic Ocean. The high concentration of diatoms gives the sediment a 'fluffy' aspect—when dry, it looks like glass wool.

(b) *Top:* Mixed radiolarian and calcareous ooze, typical of tropical oceans, especially the tropical Pacific. The mottling is due to the burrowing activities of benthic organisms (bioturbation).

Bottom: Calcareous ooze, formed of coccoliths and foraminiferal remains, between them making up about 90% of the whole.

At first sight, deep-sea sediments are little more than soft muds of various hues, from white to grey to reddish-brown (Figure 1.3). Two main kinds of deep-sea sediments can be recognized.

Terrigenous sediments are formed by weathering and erosion of land areas, and are transported to the oceans by rivers, glaciers and wind. They comprise gravels, sands, silts and clays. Biogenic (or biogenous) sediments are made up of the microscopic remains of those predominantly **planktonic** marine organisms that secrete skeletons (or tests) of calcium carbonate or silica. Figure 1.4 and Table 1.1 summarize the distribution of deep-sea sediments, and Figure 1.5 is a map of the major physiographic features of the ocean basins.

Table 1.1 Percentage of deep ocean floor covered by pelagic sediments.

Sediments	Atlantic	Pacific	Indian	World
Calcareous ooze	65.1	36.2	54.3	47.1
Pteropod ooze	2.4	0.1	—	0.6
Diatom ooze	6.7	10.1	19.9	11.6
Radiolarian ooze	—	4.6	0.5	2.6
Pelagic clays	25.8	49.0	25.3	38.1
Relative size of ocean (% of total)	23.0	53.4	23.6	100.0

QUESTION 1.1 Examine Figures 1.4 and 1.5. Is there any correlation between the distribution of calcareous and siliceous sediments and red clays, and the main physiographic features? Where is most of the terrigenous sediment to be found?

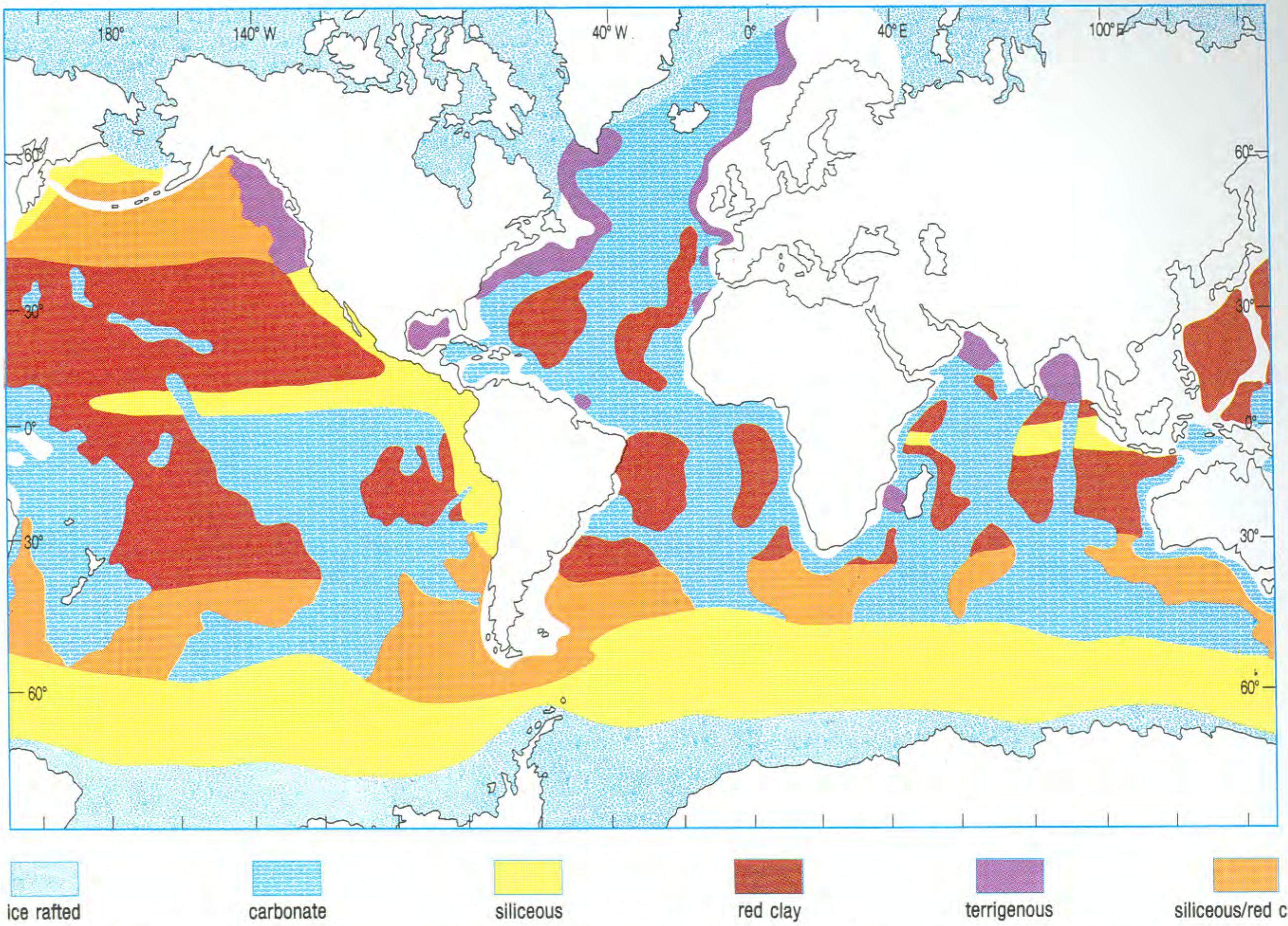


Figure 1.4 Distribution of dominant sediment types on the floor of the present-day oceans. Note that red clays are also terrigenous sediments.

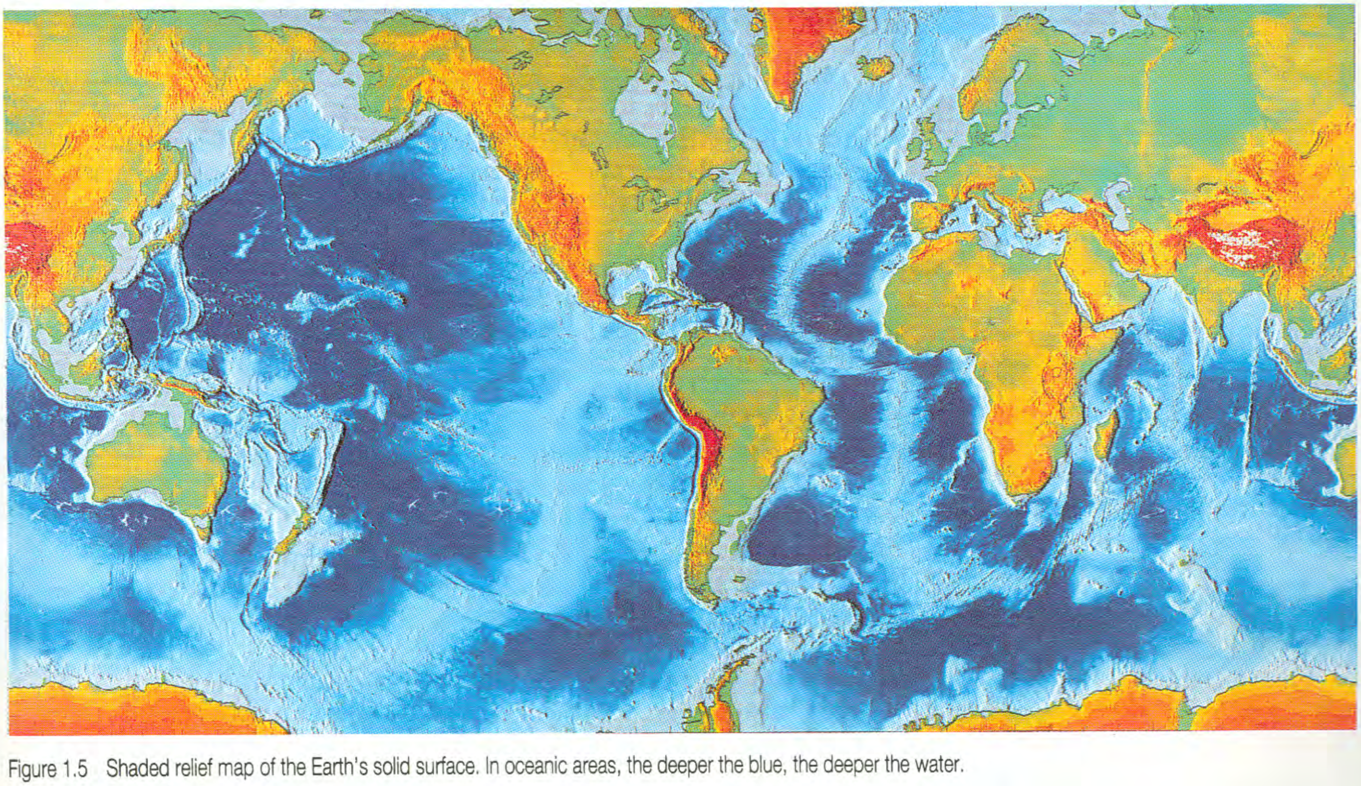


Figure 1.5 Shaded relief map of the Earth's solid surface. In oceanic areas, the deeper the blue, the deeper the water.

At this point, we must emphasize that there is no such thing as a 'pure' terrigenous or biogenic sediment. For example, terrigenous dust is widely dispersed by winds and currents; biogenic sediments always contain material of non-biological origin; and most calcareous sediments contain some siliceous material, and vice versa. Conversely, terrigenous sediments are seldom without a biogenic component, however small; and it is worth mentioning that the predominance of red clays on abyssal plains is due to an *absence* of material which elsewhere dilutes the terrigenous component, rather than to a large flux of clay particles to the sea-bed from overlying waters.

There is a convenient 'rule of thumb' for broadly classifying deep-sea sediments, known as the '30% rule'. If the sediment contains more than 30% biogenic components, it is called a calcareous or siliceous ooze (depending on which biogenic component is dominant); if it contains less than 30%, it is a red clay.

Sediments deposited in the open ocean and beyond the influence of processes along continental margins are called **pelagic sediments**. This term encompasses both terrigenous and biogenic material that accumulates slowly from dispersed suspension in the oceanic water column. It thus excludes, for example, sediments that reach the abyssal plains from turbidity currents (see Chapter 4), or the sediments round oceanic islands. Two other sources of sediment to the oceans should be mentioned here, although they generally make up rather a small proportion of the whole. Volcanic eruptions eject ash and dust into the atmosphere, and this material may be carried great distances by the wind and washed out directly into the oceans by rain. The volcanogenic component of deep-sea sediments can be significant in the vicinity of active volcanoes and after major eruptions such as that of Krakatoa in 1886. There is also a small proportion of material of extra-terrestrial origin (Figure 1.6), the remains of meteorites destroyed in their passage through the Earth's atmosphere. The rate of accumulation of this meteoritic or cosmic dust in the deep sea is estimated to be in the range 0.1 to 1mm per million years.

Figures 1.4 and 1.5 are mere snapshots in terms of geological time. The ocean basins change their shape and size at rates of centimetres per year because of **sea-floor spreading** and **plate tectonics**; and global climatic regimes fluctuate on time-scales of 10^4 to 10^7 years. The oldest ocean floor was formed about 160Ma (million years) ago, but even that seemingly vast time-span represents only about one-thirtieth of the age of the Earth as a whole. Nonetheless, during that same time-span, the Earth has seen the break up of supercontinents and a transition from an essentially ice-free planet to one with polar ice-caps. The familiar current systems of the modern ocean did not exist 160Ma ago, and the distributions of temperature and salinity—not to mention the nature and variety of most marine life forms—were quite different from what they are now.

Does that summary imply anything about what you might find if you were to drill down through the sediments in those parts of Figure 1.4 that lie well away from the ocean ridges, i.e. where the sediments are thickest?

You would still be able to recognize the broad subdivisions of sediment *type*, that is, terrigenous versus biogenic, and calcareous versus siliceous.

But those categories would almost certainly change with depth in the sequence (Figure 1.7); and secondly, within those categories, the proportions and types of non-biological components would probably be different. The species composition of the biogenic components would also be different—the more so, the deeper you drilled.

Equally important is the fact that while sediment sequences such as the one in Figure 1.7 accumulated on top of a particular piece of oceanic crust, that piece of crust will not be in the same geographical position as it was when first formed. New crust is formed at oceanic ridges and moves away from them, so that after even a few tens of millions of years it may be hundreds of kilometres from its place of origin.

Moreover, the sediments deposited on that piece of crust would no longer have their original chemical and mineralogical characteristics. Reactions between seawater and sedimentary particles begin during the comparatively short space of time taken to reach the sea-bed (rarely more than a few weeks). Reactions continue both at the sea-bed and within the growing sediment pile that accumulates as sediments continue to be deposited, because there is still plenty of seawater left in the pore spaces between individual sediment particles (see Chapter 5).

Finally, and perhaps most importantly, the record will be incomplete, because sedimentation is rarely continuous at any site. Episodes of erosion or non-deposition are common, as patterns of currents change and evolve.

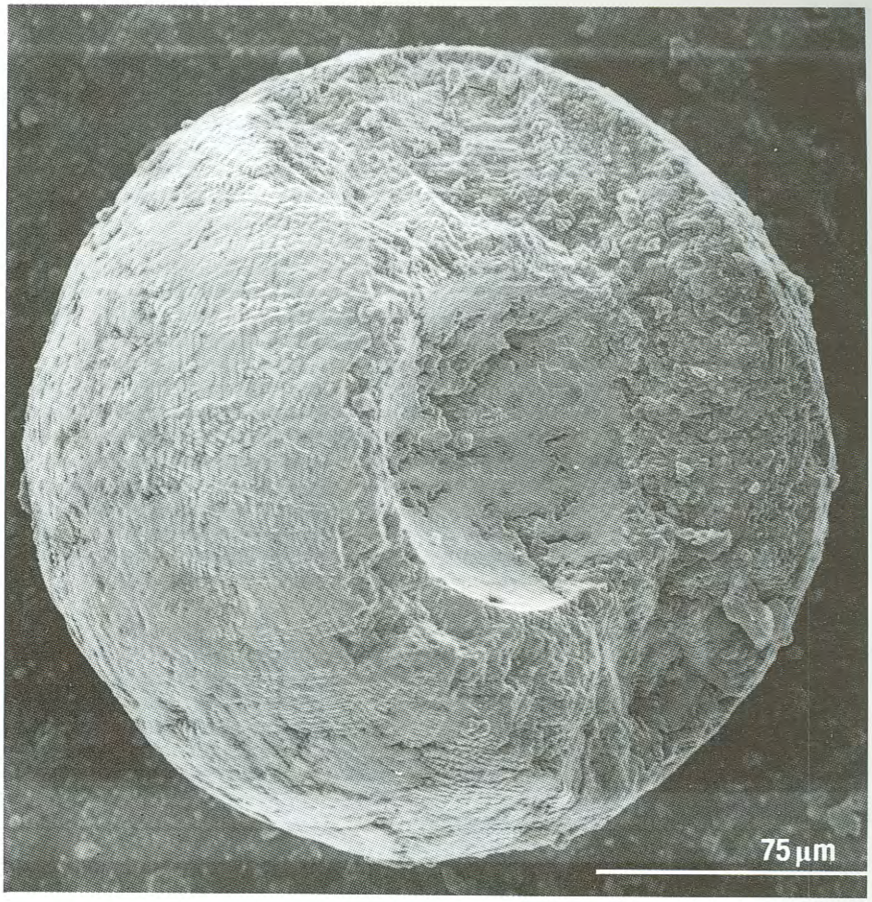


Figure 1.6 Scanning electron micrograph of a spherule of cosmic dust, typical of those found in deep-sea sediments. An iron-nickel core is surrounded by magnetite formed by oxidation during passage through the Earth's atmosphere. This spherule was found in red clay collected by HMS *Challenger* from the southern mid-Pacific. Cosmic spherules range from about 50 to 200μm in diameter and from about 0.5 to 10μg in weight. It has been estimated that 300×10^3 tonnes of cosmic (meteoritic) dust falls on the Earth's surface each year.

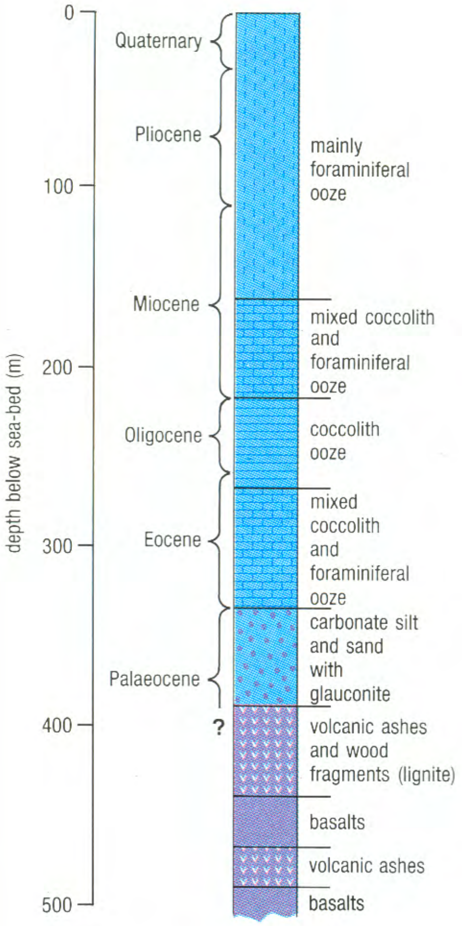


Figure 1.7 The sequence of sediments and volcanic rocks sampled at DSDP site 214 on the Ninety-East Ridge (Indian Ocean) at about 11°S, water depth 1665m. *IMPORTANT*: the sequence is presented merely to demonstrate that the type of sediment deposited at a single location may change greatly with time. You do *not* need to recognize or understand the details at this stage; they will be explained later. For geological ages, see the Appendix. DSDP = Deep Sea Drilling Project, now superseded by Ocean Drilling Programme (ODP).

1.1.1 PELAGIC BIOGENIC SEDIMENTS

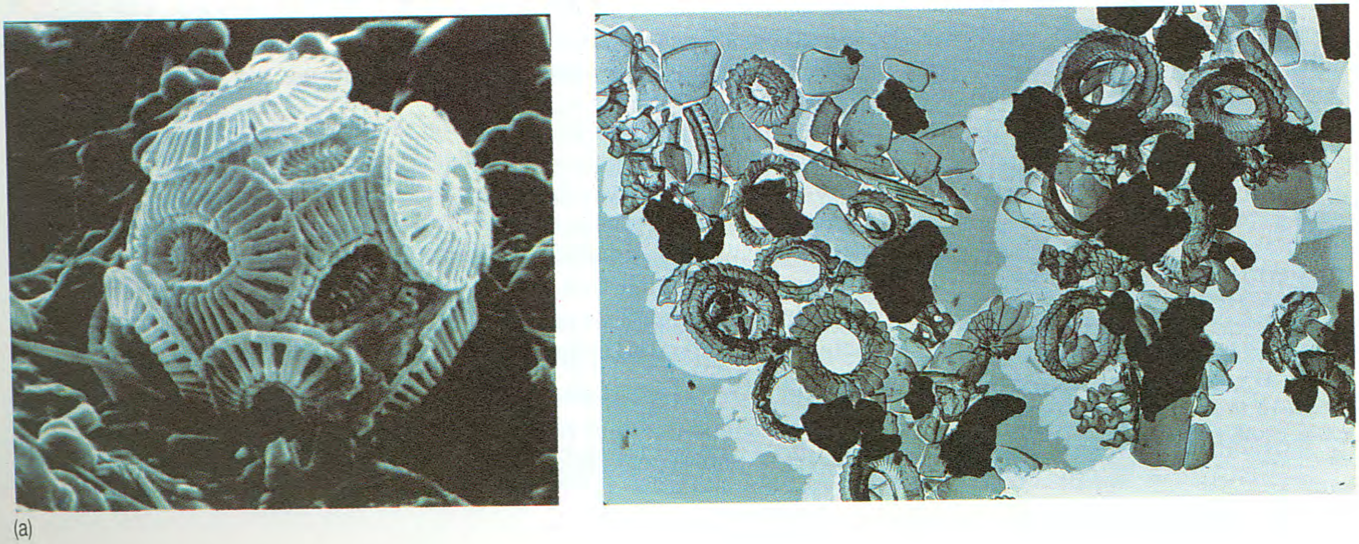
Viewed through a microscope, biogenic sediments are seen to consist of a wonderful variety of delicate and intricate structures, mostly the skeletal remains of marine phytoplankton and zooplankton. As the lifespan of most of these planktonic organisms is only about a week or two, there is a slow but continuous 'rain' of their remains down through the water column to build up successive layers of sediment. As you will see shortly, the occurrence of the remains of any one particular type of planktonic organism depends upon a number of local factors such as water chemistry and depth, and the extent of primary production in the surface ocean waters. Because of this, the presence of these microfossils in ancient deep-sea sediments can be used to determine what the water depths and surface productivity were like during the geological past.

Carbonate sediments are composed principally of the skeletal remains of **coccolithophores**, **Foraminifera** and **pteropods**. Siliceous remains come mostly from **diatoms** and **Radiolaria**. The hard parts of these organisms vary a great deal in size, shape and chemical stability, and these factors control their preservation potential in deep-sea sediments.

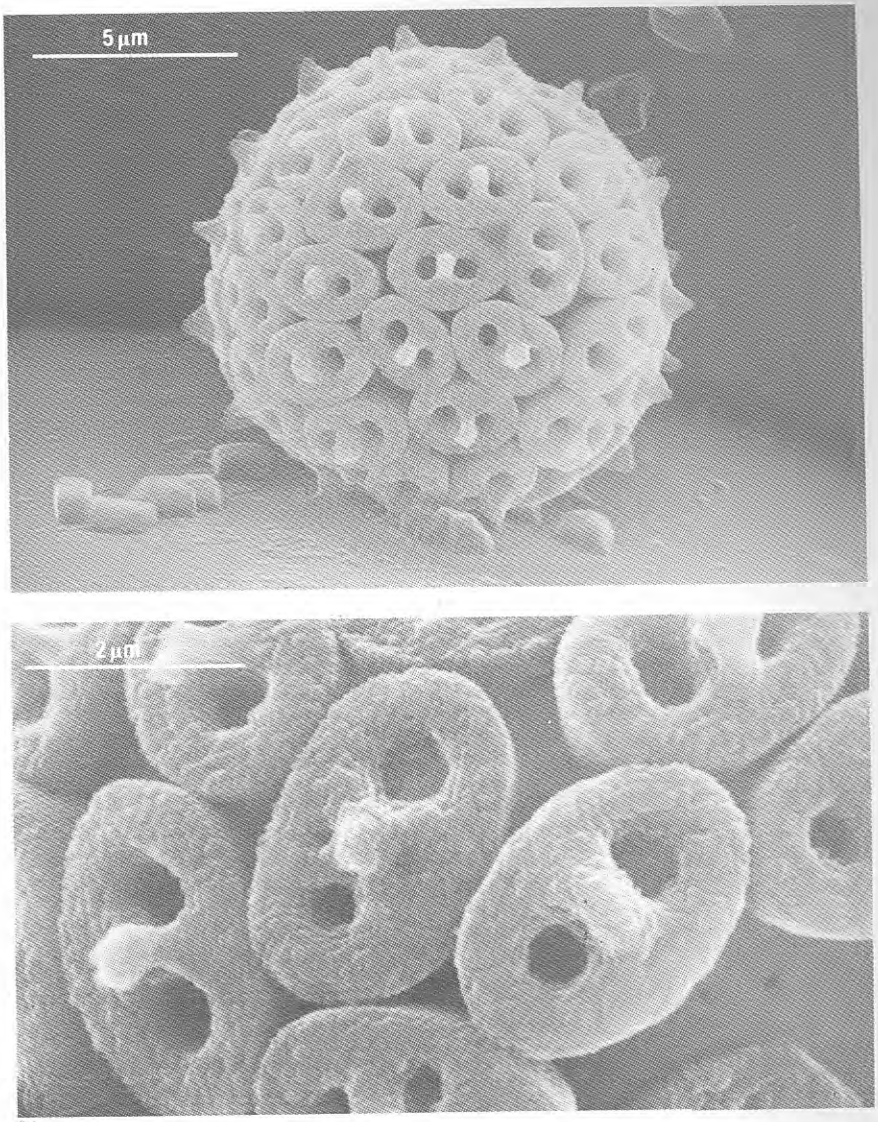
Coccoliths are minute plates of **calcite** (the more common form of calcium carbonate, $CaCO_3$), usually less than 10 μm in size ($\mu m = $ micrometre (micron) $= 10^{-6}$ m), with which the phytoplanktonic (algal) coccolithophores envelop themselves to form coccospheres (Figure 1.8). When the algal cell dies, the coccosphere disintegrates easily, releasing the individual plates—the coccoliths—into suspension. Thus, it is the coccoliths rather than the whole coccospheres that are preserved in sediments. Plates are also shed from the coccospheres as the algae grow, so coccolithophores contribute to deep-sea carbonate sediments before the death of the organism, as well as afterwards. Each coccolith has an organic membranous covering which inhibits dissolution of the calcite and enhances preservation. The white cliffs of Dover are formed of chalk that is mostly composed of coccoliths.

Because of their exceptionally small size, coccoliths are referred to as **nanofossils** (from the Greek word 'nanos', meaning dwarf) and carbonate sediments particularly rich in coccoliths are known as **nanofossil oozes** (or simply as coccolith oozes or nano-oozes).

Figure 1.8 Examples of coccolithophores.
(a) *Coccolithus huxleyi*. Both complete and disaggregated coccospheres; coccosphere diameter *c.* 6 μm.



(a)

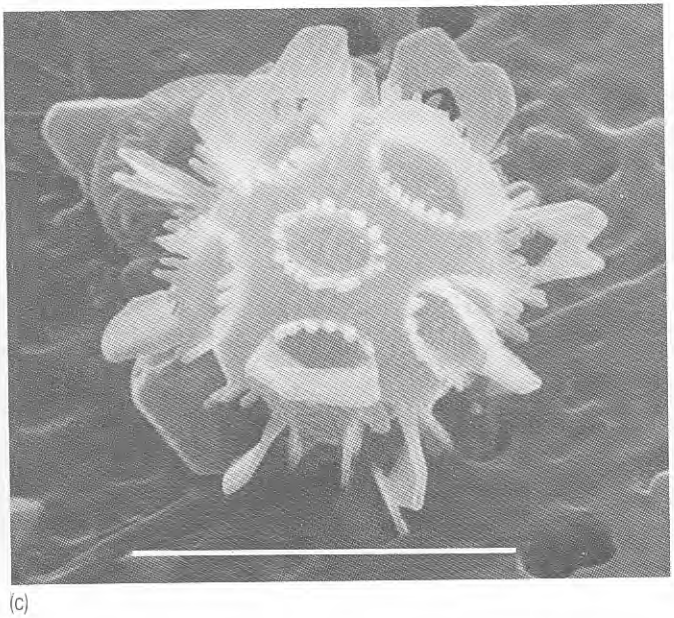


(b)

Figure 1.8 (b) *Homozygosphaera ponticulifera*. Both complete and disaggregated coccospheres; scale bars 5 µm and 2 µm.

(c) *Gephyrocapsa ornata*. Complete coccosphere; scale bar 5 µm.

(d) *Scyphosphaera apsteinii*. Complete coccosphere; scale bar 20 µm.



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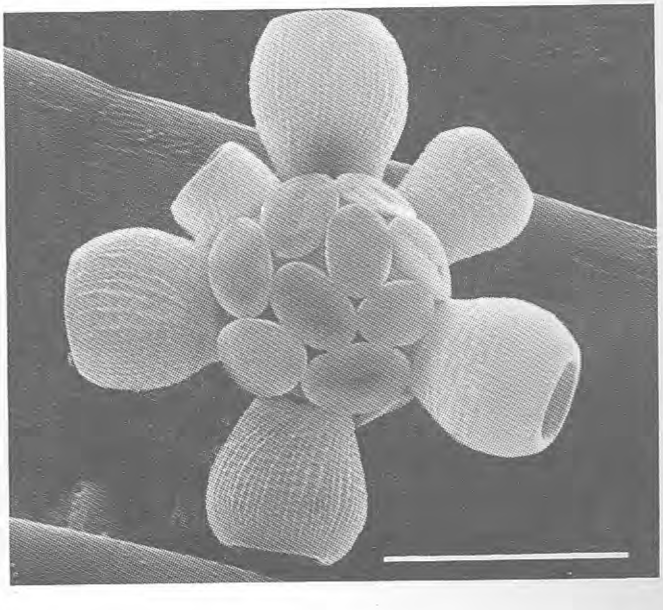
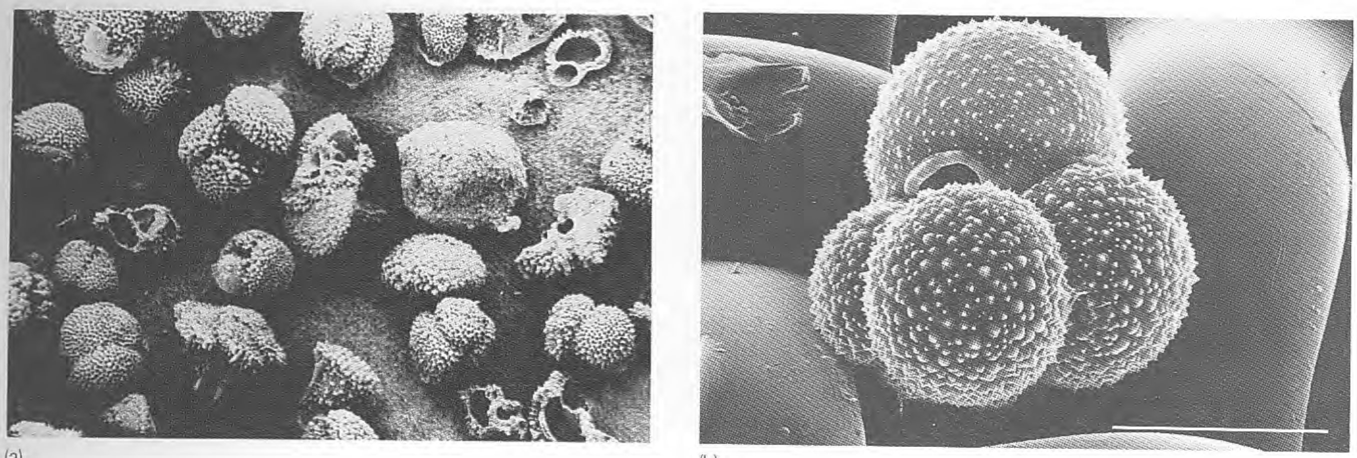



(a)

(b)

Figure 1.9 (a) Foraminifera from washed calcareous ooze. See also Figure 1.2(b).

(b) *Globigerina* sp. Scale bar 50 μm. The background is the mesh of the sampling net.

Foraminifera produce calcite exoskeletons, or tests, in the size range 30 μm to 1 mm (Figure 1.9).

Most zooplanktonic Foraminifera live in the top 1000 m of the water column. Surface and near-surface species tend to have a more spiny shape, which increases the ratio of surface area to volume and thereby aids a floating mode of life, as well as deterring predators. However, an abundance of spines also increases the surface area of a test, making it more prone to dissolution after death. Deeper-water species, which usually lack spines, are more commonly preserved. Because of their relatively large size, abundant Foraminifera give a sandy texture to the pelagic sediments in which they occur. There are also both shallow-water and deep-water forms of bottom-living or **benthic** Foraminifera, and in some regions these can be abundant.

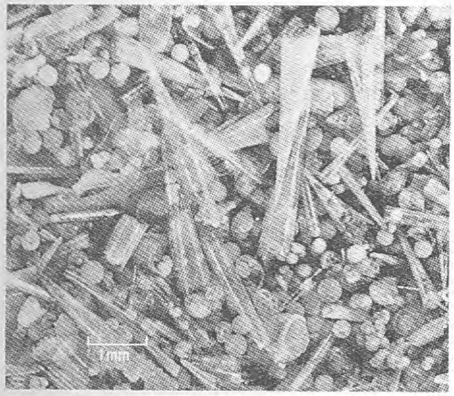


Figure 1.10 Pteropods. The larger shells are about 3mm long. See also Figure 1.2(c).

Pteropods are planktonic gastropod molluscs with thin shells up to a centimetre long (Figure 1.10). At the present day, most species are restricted to tropical and sub-tropical oceanic areas. Their shells are composed of **aragonite** (a variety of calcium carbonate, $CaCO_3$, that is more soluble than calcite), so they are more easily dissolved than calcitic plankton shells and are not found in ocean sediments where water depths exceed about 2–3 km.

Diatoms are unicellular algae ranging in size from a few μm to around 200 μm (Figure 1.11). They secrete shells of amorphous hydrated silica, the formula of which is commonly written as SiO_2nH_2O (and it is sometimes called opaline silica or opal). For brevity, we shall use SiO_2. Both planktonic and shallow-water benthic diatoms occur, but the planktonic species have thinner tests and are more prone to dissolution. The remains of planktonic diatoms often dominate the siliceous sediments found at high latitudes (Figure 1.4).

QUESTION 1.2 (a) Why are benthic diatoms restricted to shallow water?

(b) Can you suggest the maximum depth at which you might expect to encounter benthic diatoms?

Radiolaria are quite large zooplanktonic organisms (Figure 1.12), usually between 50 μm and 300 μm or more in size. They also have skeletons of silica and are the dominant biogenic component of siliceous sediments

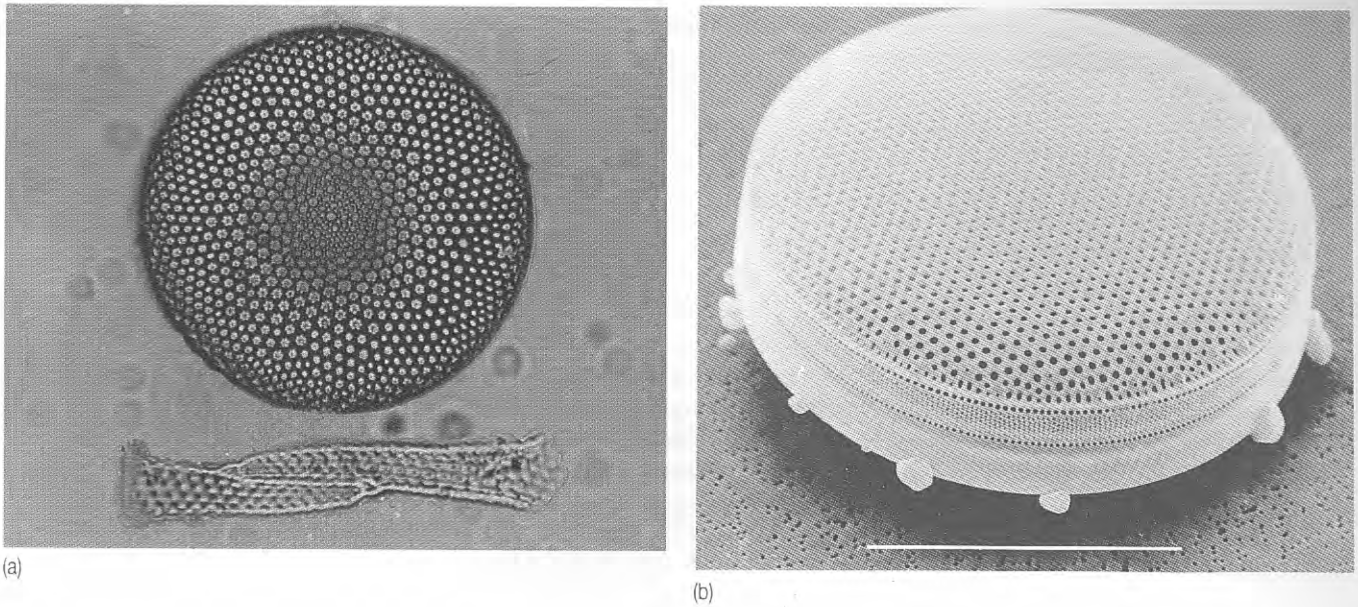


Figure 1.11 (a) Discoid (centric) and spindle-shaped diatoms. Width of field about 50 μm.

(b) Centric diatom. Scale bar 20 μm.

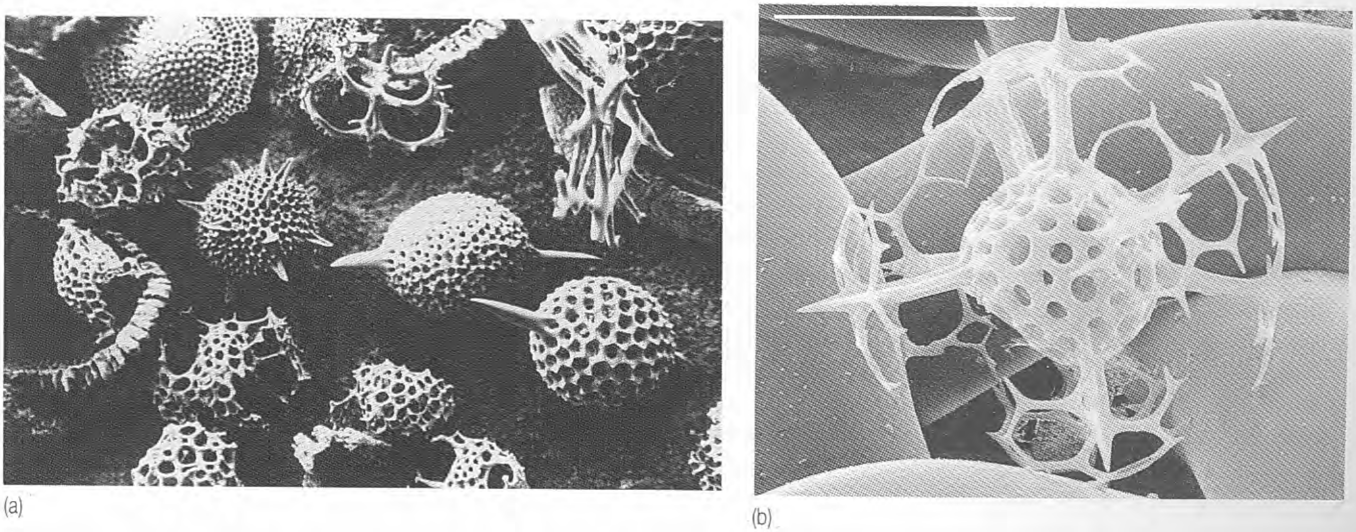
found at low latitudes. As with the Foraminifera, both surface and deeper-water species occur.

QUESTION 1.3 By analogy with Foraminifera, how would you expect the shape and preservation potential of surface forms of Radiolaria to differ from those of deeper-water forms?

It is important always to keep in mind that the distribution of biogenic sediments (Figure 1.4) is determined partly by the extent of biological productivity of the plankton in surface waters; and partly by the extent to which the skeletal remains are dissolved in the water column and at the sea-bed—and that in turn depends upon the chemical properties of seawater (see Chapter 3).

Figure 1.12 (a) Radiolaria from washed siliceous ooze. Individual tests *c.* 100 μm across. See also Figure 1.2(a).

(b) *Hexacontium* sp. Scale bar 50 μm. The background is the mesh of the sampling net.



1.1.2 PELAGIC TERRIGENOUS SEDIMENTS

Nearly all terrigenous sediments in the pelagic environment are composed of material of the smallest grain sizes. There are two ways in which coarser-grained material can reach the pelagic environment. One is via turbidity currents and other gravity flows (see Chapter 4). The other way is through ice-rafting, that is, glacial material being shed by icebergs drifting into the open ocean and melting. Ice-rafted boulders, pebbles and sand may be found among pelagic sediments up to several hundred kilometres from the glaciers or ice-shelves from which they originally 'calved'.

The wind is obviously a powerful means of transporting fine material directly to the open oceans. Indeed, it is important to emphasize that pelagic clays deposited in the open oceans are predominantly of wind-blown (aeolian) origin and mostly less than about 20μm in size. The regions most likely to generate wind-blown dusts are the low-latitude belts influenced by persistent Trade Winds and low rainfall. The total amount of wind-blown dust delivered to the oceans annually is of the order of 10^8 tonnes. This is very small compared with the sediment load supplied by rivers, which is about 1.5×10^{10} tonnes per year. Some river-borne sediment does reach the abyssal plains (see Chapter 4) but nearly all of it is deposited along continental margins; that includes clay minerals which are largely removed from suspension by **flocculation** in river mouths and estuaries. So, the proportion of river-borne terrigenous sediment in pelagic clays can be reduced relative to the aeolian contribution.

The principal components of wind-blown dusts are quartz and clay minerals, and these are being continually supplied to the oceans. More intermittent and much more spectacular is the supply of volcanogenic material to the pelagic environment though, on a global scale, the proportion of such material in deep-sea sediments is small (Section 1.1). Major eruptions can eject large quantities of volcanic ash and dust to heights of 15 to 50km, where the smallest particles, 1μm or less, can remain suspended for many months. During that time, they can be carried several times round the world by high altitude winds, causing unusual weather conditions and spectacular sunsets. Material between about 1μm and 20μm in size will rarely be projected to heights above 10km, and is deposited in a matter of days or weeks and within several hundred to a few thousand kilometres from the eruption. The result is distinctive layers of volcanic ash, which can be useful in the correlation of pelagic sediment sequences from widely separated locations.

There is a growing number of wind-borne anthropogenic contributions to the deep sea: some benign, others less so. They include dust from power stations and cement works (Figure 1.13); particles of more persistent plastics; PCBs (polychlorinated biphenyls from the plastics and electrical industries); lead compounds (mainly from motor vehicles); radionuclides; and various products from waste incineration at sea. The production of these materials is limited to the past hundred years or less, so their appearance in sediments can be used as a time-marker and in some cases also as a tracer for the movement of material through the oceans, on its way to the sea-bed.

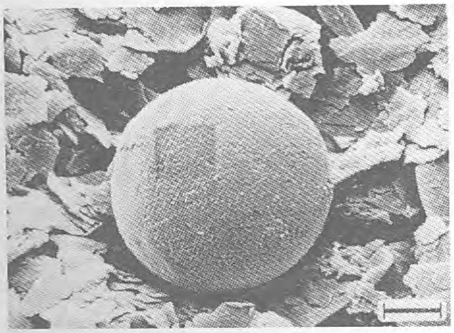


Figure 1.13 Scanning electron micrograph of a fly-ash particle embedded in biogenic debris and clay minerals, collected in a sediment trap on the floor of the Sargasso Sea. Scale bar = 3μm.

The pelagic clays or so-called red clays (Figure 1.3) are found over large areas of the deep ocean floor, especially in the Pacific (Figure 1.4). They comprise a number of different types of clay minerals and, as we said earlier, owe their existence as a recognizable sediment type to the *absence* of other sediment components rather than to an abundance of clay minerals as such. Thus, pelagic clays are least diluted in the deepest parts of the ocean basins, and away from areas of high surface productivity. In these regions, not only is there a lack of biogenic material sinking from the surface, but also the chemistry of the bottom waters is such that any calcareous shell debris is dissolved (see Chapter 3). These regions are also well away from continental margins and so the clays are less diluted by other terrigenous sediments. Their colour is actually brown, rather than red, and results from oxidation of iron in the sediments, which have very low sedimentation rates and experience prolonged exposure to oxygenated bottom waters on the sea-floor.

Clay minerals are hydrated aluminosilicate minerals that occur as thin flakes, generally less than 2µm across. There are four main types of clay minerals in pelagic sediments: kaolinite, chlorite, illite and montmorillonite, each of which is formed in a different weathering environment. Their relative proportions in a pelagic clay vary according to the prevailing climatic and geological conditions in the source region and along the transport pathways, and according to the mixing processes that occur in the oceans.

Kaolinite is formed by the extreme chemical weathering of silicate minerals, especially feldspars, and is most abundant in low latitudes.

Chlorite occurs in both igneous and metamorphic rocks of the continents, but is destroyed by the chemical weathering that predominates in low latitudes. In general, therefore, chlorite is abundant in pelagic clays in high latitude environments, where physical weathering predominates and chlorite is released to the oceans in an unaltered state.

Illite is the most widespread clay mineral, but is more abundant in the Northern Hemisphere, where it may contribute up to 70% of the clay minerals in a sediment. Illites form under a variety of conditions and are not characteristic of any particular latitude belt; so, whether they dominate the sediment or not depends on the degree of dilution by other clay minerals.

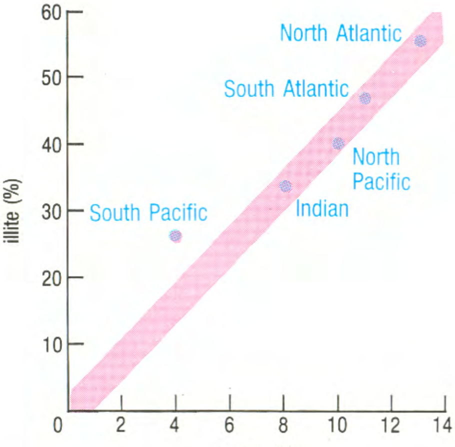


Figure 1.14 The relationship between the average concentrations of quartz and illite in the <2µm size fractions of sediments of the major ocean basins.

QUESTION 1.4 How well does Figure 1.14 confirm that last statement?

Montmorillonite is an alteration product of volcanogenic material, both in continental areas and in the marine environment. Much of the montmorillonite in deep-sea sediments is produced by the 'weathering' of volcanic ash actually on the sea-floor. Thus, strictly speaking, it is not always a terrigenous mineral, because some of it is formed *in situ* (see Section 5.2). It has been estimated that a layer of volcanic ash in a deep-sea sediment sequence can be totally altered to montmorillonite-rich pelagic clay in about 20Ma.

1.2 SEDIMENTS AND SEAWATER

The oceans receive a great variety of both solid and dissolved material from different sources. Deep-sea sediments are dominated by three types of components: calcium carbonate, silica and clay minerals (Figure 1.4). Relative amounts of these components in the sediments (Table 1.1) are significantly different from what they are in the particulate material that sinks from the surface. This is dominated by organic matter making up the body tissues of marine plants and animals, only *some* of which form skeletons of calcium carbonate or silica. Hardly any of the organic matter survives to reach the sea-bed—most of it is consumed and recycled in the top kilometre of the water column. Varying proportions of the skeletal materials are also dissolved before they can accumulate in the sediments.

By contrast, clay minerals and other inorganic solid products of terrestrial weathering are quantitatively almost insignificant in surface waters. However, they are relatively little affected by their passage through the water column, which is why they are so abundant in deep-sea sediments (Table 1.1).

We have seen that calcium carbonate and silica are being extracted from solution in seawater to form skeletons of plants and animals. But what about the other dissolved constituents? There is no evidence that the oceans are getting saltier with time, so these other constituents are somehow also being processed in the oceans and removed from solution. In the next Chapter, we examine the processes that cycle dissolved constituents within the main body of the oceans and ultimately remove them to the sediments.

1.3 SUMMARY OF CHAPTER 1

1 Deep-sea sediments can be classified broadly as terrigenous (land-derived) and biogenic (formed as the result of biological activity), with minor volcanogenic and cosmic contributions. Pelagic sediments include all those sediments deposited in the deep ocean basins beyond the influence of continental margin processes. Deep-sea sediments comprise material from more than one source. Factors such as submarine topography and climatic patterns influence the type of sediment that accumulates in a particular region.

2 The present-day distribution of deep-sea sediments is a reflection only of present-day climatic and current patterns and of the present configuration of ocean basins. The type of sediment deposited on a particular piece of ocean floor changes with time because of the way in which ocean basins, currents and climate change. The sequence of sediments recovered in a sediment core preserves a partial record of the changes that have occurred in the ocean above it.

3 Pelagic biogenic sediments are composed mostly of the remains of calcareous (carbonate) and siliceous planktonic organisms, principally coccolithophores, Foraminifera, pteropods, diatoms and Radiolaria. The preservation of these depends upon a number of factors such as water depth and chemistry, the shape of the skeletal remains, and the presence or absence of an organic membrane.

4 Wind-blown dusts comprise quartz and clay minerals and volcanogenic components. Coarse debris occurs in pelagic sediments only in high latitudes, deposited by ice-rafting. Pelagic clays comprise clay minerals derived from a number of different sources. They predominate in the deepest parts of the ocean basins where they are not diluted by biogenic material. Four main types of clay minerals are recognized, each characteristic of particular weathering regimes: kaolinite, chlorite, illite, and montmorillonite. Aeolian inputs are a major contribution to pelagic clays in many regions.

Now try the following questions to consolidate your understanding of this Chapter.

QUESTION 1.5 Suggest two reasons why sediments should be thicker near to continental margins than near mid-oceanic ridges.

QUESTION 1.6 It has been established by careful analysis that the ratio of the concentration of calcium to the total salinity of seawater is greater in deep than in surface water. Would you expect the ratio of dissolved silica (SiO_2) to total salinity to vary in the same way?

QUESTION 1.7 Explain whether you would expect to find chlorite or kaolinite in sediments: (a) of the equatorial Atlantic, (b) round Antarctica. (c) Why should illite be more common in sediments of the Northern than the Southern Hemisphere?