10 Deep-Sea Sediments: Patterns and Processes

10.1 Legacies of HMS *Challenger* and Other Pioneer Vessels

10.1.1 Challenger, Meteor, and Albatross Pioneers and Vema and Glomar Challenger

Deep-sea deposits were first explored in a comprehensive fashion during the British *Challenger* Expedition (1872–1876). Many thousands of samples were studied by the Scottish naturalist John Murray (1841–1914), participant of the expedition and chief pioneer of deep-sea geology. He and his coworker, the Belgian geologist A.F. Renard (1842–1903), published a weighty report on the results, a tome that laid the foundation for later research in the field of deep-sea geology, with emphasis on sediments. The first distinct step beyond Murray's *Challenger*-based studies was taken almost half a century later by the German *Meteor* Expedition (1927–1929), a cruise that took regularly spaced short cores in the central Atlantic. A new branch of oceanography started with the recovery of long cores (7 m, typically) on a global scale by the Swedish *Albatross* Expedition (1947–1949), that is, Pleistocene paleoceanography. It started the revolution of our understanding of climate and ice ages.

The pioneers associated with post-*Challenger* developments were the German geologist Wolfgang Schott of the *Meteor* Expedition (then in his twenties) and the Swedish radiochemist and physicist Hans Pettersson (1888–1966), leader of the *Albatross* Expedition (then in his fifties). Gustaf Arrhenius (now a retired professor at S.I.O.), a geochemist and a young member of the Albatross Expedition, described the carbonate cycles from the equatorial Pacific that became crucial in documenting the ice ages.

An early culmination and first flowering of the approach of retrieving and studying deep-sea cores to elucidate the geologic history of Earth were manifested in the work of the marine geophysicist Maurice Ewing (1906–1974), founder of the Lamont Observatory at Columbia University, and many of his associates. Maurice Ewing insisted on gathering large numbers of long cores from the *RV Vema*; cores used later especially by the CLIMAP members. One great step forward came in 1968 with the first leg of the Deep Sea Drilling Project (DSDP) and the *Glomar Challenger*, which drilled for samples of Cenozoic and Cretaceous ocean history. Maurice Ewing led the first leg. S.I.O.'s geochemist M.N.A. Peterson (1929–1995) wrote much of the blueprint for early Deep Sea Drilling Legs.

Perhaps the single most important finding of John Murray was that the non-clay deep-sea sediments everywhere largely consist of calcareous shell material, supplied by plankton organisms. It took decades after the *Challenger* Expedition to get a good estimate of the rate of shell supply to the seafloor. From comparing the abundance of shelled plankton in the surface-near waters (where most of the growth occurs) with the rate of sediment supplied, one could make the first trustworthy estimates about the rate of production of the shelled plankton, the chief source of Earth's sediment cover. That sedimentation rates are exceedingly slow for these deposits (ca. 1 cm/1000 years or slightly more) was first found by W. Schott based on *Meteor* cores, using (the very rough) varve dates for the end of the last glacial period obtained by the Dutch geologist de Geer (1858–1943) from counting layers in lake sediments near icy areas and assuming they represent annual layers.

Regarding fine-grained sediments, a major problem arose for Murray and Renard. Available optical equipment was good down to fine sand. X-rays had not been discovered yet. Naturally, not having modern equipment for the study of clay-sized particles, Murray was mainly concerned with coarse particles (sand size and up to small pebble size for certain pelagic mollusk shells). For this reason (and others), he chose names such as *pteropod ooze* and *Globigerina ooze* for the sediment that covers the shallower half of the deep seafloor. *Globigerina* is a common genus of foraminifers; for tropical spine forms, the term *Globigerinoides* is more appropriate (Fig. 10.1 , left). In any case, the major discovery was that the sand-sized shells were remains from plankton, not benthos.

E. Seibold, W. Berger, *The Sea Floor*, Springer Textbooks in Earth Sciences, Geography and Environment,
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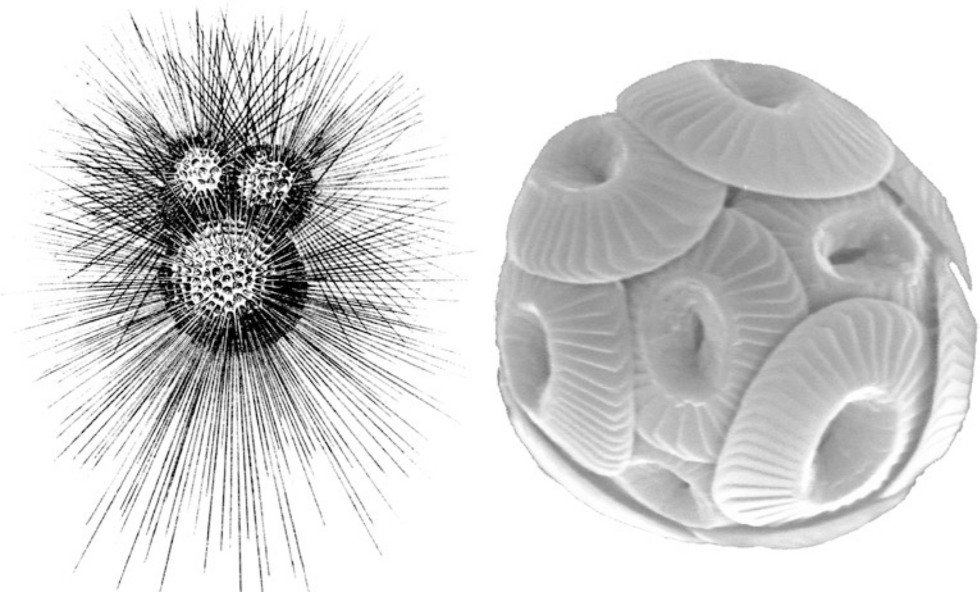

Fig. 10.1 Dominant sediment particles in calcareous deep-sea sediments. *Left*: Planktonic foraminifer (in the water); from H. Brady, Challenger Expedition. *Right*: coccolithophore, SEM graph (Courtesy R. Norris; images not to scale, the coccosphere being roughly ten times smaller than the foraminifer)

Today, we prefer the more general term *carbonate ooze* over "Globigerina ooze" considering that much of the plankton-derived material over most of the seafloor consists not of shells of foraminifers (including the genus *Globigerina*) but of the remains of minute calcareous algae, the *coccolithophores.* The skeletal elements of these minute microbes are studied largely with the aid of scanning electron microscopes (*SEM*), which became available well after Murray and Renard wrote their book. Coccolithophores (Fig. 10.1, right) are plankton organisms that shed the exceedingly small coccoliths, fine-silt fossils that are ubiquitous in calcareous marine deposits (*nannofossils*). Cenozoic nannofossils, like most fossils, largely consist of extinct forms. They are very useful in biostratigraphy, a fact established largely by the US geologist M.N. Bramlette (1896–1977) and his associates. The nannofossils include abundant *Discoasters*, last common in the Pliocene, several million years ago.

10.1.2 Calcareous Ooze and "Red Clay": Discovery of the "CCD" (Carbonate Compensation Depth)

Accepting that the discovery of the plankton connection of fossils on the deep seafloor was a major pioneering feat (Figs. 8.16 and 10.1), what were other chief insights that emerged and became available through the labors of Murray and Renard? Two major results come to mind as being central to the understanding of deep-sea deposits (i.e., deposits below the shelf edge). The foremost one concerns the major sedimentary boundary between the calcareous *ooze* on the upper half of the deep seafloor (buff to cream-colored fossil assemblages of microscopic plankton) and the *Red Clay* in the lower half (fine-grained reddish-brown residue after shells have been removed by dissolution) (Fig. 10.2).

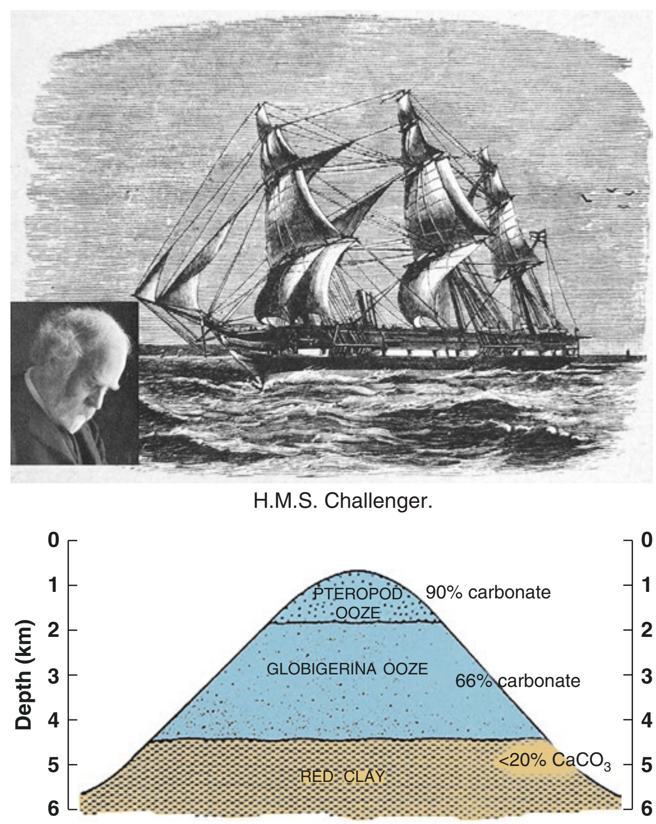

Fig. 10.2 A major result of the nineteenth-century *Challenger* Expedition: the CCD. Dominant types of deep-sea sediment are carbonate "oozes" and "Red Clay" (the residue after dissolution of carbonate ooze, commonly at depths well below 4 km). The calcareous oozes cover almost half of the deep seafloor (i.e., about one-third of the solid Earth); Red Clay covers about 40% of the deep seafloor (i.e., only slightly less than calcareous ooze). The remainder is largely diatom ooze, that is, it is highly siliceous. Note the typical carbonate contents, decreasing with depth (Images after John Murray, naturalist of the Expedition: inset (portrait of Murray well after the expedition) from Murray and Hjort, 1912)

10.1.3 Deep-Sea Deposits of Green Versus Blue Ocean

The third most important insight was the discovery that sediments surrounding the continents are not at all like those of the deep sea but are dominated by terrigenous contributions (i.e., weathering products from continents) rather than plankton shells and have abundant benthic microfossil fossil remains (Fig. 10.3). Also, they contain enormous amounts of organic material produced by the coastal ocean. Although Murray could not know that, sedimentation rates are typically almost ten times higher on continental slopes than in the deep sea (ca. 10 cm per millennium rather than the roughly 1–2 cm/1000y for calcareous ooze). What Murray could easily see is that many of the continent-near deposits have a greenish hue (signifying oxygen shortage) rather than the brown and buff colors that dominate modern deep-sea deposits away from continents (signifying availability of plenty of oxygen). The high-production aspect of the deposits surrounding the continents also includes a rich assortment of siliceous materials, notably the shells of robust diatoms. Also, the high supply of organic matter in the coastal ocean results in increased dissolution of carbonate from the acidification of interstitial waters that comes with the generating of carbon dioxide from the oxidation of organic matter.

Thus, the Challenger Expedition established and documented the major features of sedimentation on the deep seafloor. However, there remained plenty to discover for the pioneers that followed. For example, the *Meteor* Expedition established that the main facies boundary (the CCD) tends to be associated with a major quasi-horizontal water mass boundary in the South Atlantic. The *Albatross* Expedition discovered that carbonate deposition varies with time in the ice ages and that the variation is cyclic. Subsequent coring by major oceanographic institutions (led by Lamont's Research Vessel Vema) established the cyclicity of ice age history in some detail, in deep-sea sediments. Finally, deep-sea drilling revealed that sediment patterns change fundamentally through geologic time ending up entirely different in the Neogene from patterns that dominated in the late Cretaceous.

Obviously the main dichotomy of deep-sea sediments was known to John Murray (Fig. 10.2). He realized that carbonate dissolution was the least intense on the shallowest portions of the seafloor, where the delicate aragonitic shells of pteropods could be preserved. The tiny mollusks are also known as "sea butterflies" in popular language, borrowing the name from the large-winged insects, that is, terrestrial arthropods, and members of another phylum. Pteropods are part of the plankton; their shells are readily seen in deep-towed nets, dissolving while falling through deep waters off California. Given their vulnerability, pteropods are thought to be the first conspicuous victims to suffer from future acidification.

10.2 Inventory and Overview

10.2.1 Sediment Types (Facies) and Distributional Patterns

Murray's simple classification scheme for deep-sea deposits is pretty much still in use after more than a century of being formally introduced, although categories are somewhat more detailed (Table 10.1).

Fig. 10.3 Ooze on the open ocean floor versus glauconitic mud on continental slope (From Murray and Renard). *Left*: well-preserved calcareous ooze (note the delicate aragonitic cone-shaped pteropod shells at the far left). *Middle*: foraminiferal ooze. *Right*: continental slope sediment; note diversity of organisms. Also note the shiny *dark-green* glauconite fill in many benthic foraminifer shells

Table 10.1 Classification of deep-sea sediments

I. (Eu-) pelagic deposits (oozes and clays)
<25% of fraction >5 μm is of terrigenic, volcanogenic, and/or neritic (shelf) origin
Median grain size is <5 μm (excepting authigenic minerals and pelagic high-sea organisms):
A. Pelagic clays: $CaCO_3$ and siliceous fossils <30%
1. $CaCO_3$ 1–10% – (slightly) calcareous clay
2. $CaCO_3$ 10–30% – very calcareous clay (or marl)
3. Siliceous fossils 1–10% – (slightly) siliceous clay
4. Siliceous fossils 10–30% – very siliceous clay
B. Oozes: $CaCO_3$ or siliceous fossils >30%
1. $CaCO_3$ > 30% < 2/3 $CaCO_3$ marl ooze; >2/3 $CaCO_3$ chalk ooze
2. $CaCO_3$ < 30% >30% siliceous fossils: diatom or radiolarian ooze
II. Hemipelagic deposits (muds)
>25% of fraction >5 μm is of terrigenous, volcanogenic, and/or neritic (shelf) origin
Median grain size is >5 μm (excepting authigenic minerals and pelagic organisms):
A. Calcareous muds: $CaCO_3$ > 30%
1. <2/3 $CaCO_3$ – marl mud, >2/3 $CaCO_3$ chalk mud
2. Skeletal $CaCO_3$ > 30% – foram ~, nanno ~, coquina mud
B. Terrigenous and other muds: $CaCO_3$ < 30%, quartz, feldspar, or mica dominant
Prefixes: quartzose, arkosic, micaceous
Volcanogenic muds: $CaCO_3$, <30%, ash, palagonite (altered volcanic matter), etc. dominant:
Appropriate prefixes, Diatom rich: siliceous mud
III. Various special deposits
1. Cretaceous carbonate-sapropelite cycles
2. Back (carbonaceous) clay and mud: sapropelites (e.g., Black Sea)
3. Silicified clay stones and mudstones: chert (largely pre-Neogene)
4. Pre-Neogene limestones

W.H.B. 1974 in C.A. Burke and C.L. Drake (eds.) *The Geology of Continental Margins*. Springer, Heidelberg and Berlin. From lists in DSDP and ODP

The main types of sediment were already known at the time of the *Challenger* Expedition, from earlier scientific voyages, but without the benefit of formal and detailed classification. The chief vertical contrast, as mentioned, is between *calcareous ooze* and *pelagic clay* (Fig. 10.2). The main contrast in regard to distance from land in essence is between pelagic deposits (oozes and clays) and hemipelagic ones (mud) (Fig. 10.3). These muds have some of the same ingredients as the pelagic sediments (clay-sized dust and volcanic ash, foraminifer shells, coccoliths, radiolarian skeletons, diatom frustules) but also bear significant indicators of high production and large admixtures of shelf-derived sediment and continental material. The list given in Table 10.1 reflects the main constituents recognized in optical microscopes and analyzed by X-ray methods (since the 1940s) and by SEM (since ca. 1970). It is abbreviated and may vary somewhat depending on author.

The tripartite nature of the traditional classification (ooze, clay, mud) is readily appreciated when contemplating overall distribution patterns (Fig. 10.4).

The tripartite nature of the deep-sea sediments readily lends itself to a schematic representation in a two-dimensional graph (the three main categories being products of depth-dependent change of carbonate of and of distance from land). The one shown (after one by the Swedish marine geologist Eric Olausson, 1923–2010) refers to the eastern central Pacific (Fig. 10.5).

On the seafloor of the world ocean, the carbonate-clay dichotomy dominates (with the clay facies marking abyssal depths), while the high-production slope sediments are typically confined to a relatively narrow band around the continents. In some places (Gulf of Alaska, Bay of Bengal , off south eastern Canada, off north western Africa), turbidite deposits have extended the mud facies into the deep sea (marked "m" in Fig. 10.4).

10.2.2 Biogenous Sediments Dominate

The bulk of the deep-sea deposits consists of biogenous sediments, notably plankton shells (Fig. 10.6). About one half of the seafloor is covered by *oozes*, that is, sediments formed of various types of plankton remains (chiefly coccoliths (ca. 5–30 μm)), foraminifer shells (ca. 50–500 μm), diatom remains (ca. 5–50 μm), and radiolarian skeletons (ca. 40–150 μm).

A few hundred meters below the sea surface, there are but few shells found on living organisms. Instead, we find shell debris on the way to the seafloor. The sinking shells, being largely made of soluble mineral matter, dissolve on the way down and on the seafloor itself. Thus, what one finds in the sediment is a selection of the more robust forms for much of the plankton. Calcareous matter is especially vulnerable to dissolution at great depth. Below a critical depth, called the *carbonate compensation depth* or *CCD*, calcareous particles are largely removed, and we obtain Red Clay, precisely as depicted by J. Murray and A.F. Renard (Fig. 10.2).

10.2.3 On the Striking Difference in Sedimentation in the Pacific and Atlantic

When comparing the sediment cover of the seafloor in the Pacific and Atlantic, one finds that the deep Atlantic seafloor preferentially accumulates carbonate, especially in the northern hemisphere, and the Pacific seafloor has more silica. We shall see (when discussing drilling results in the Neogene, in

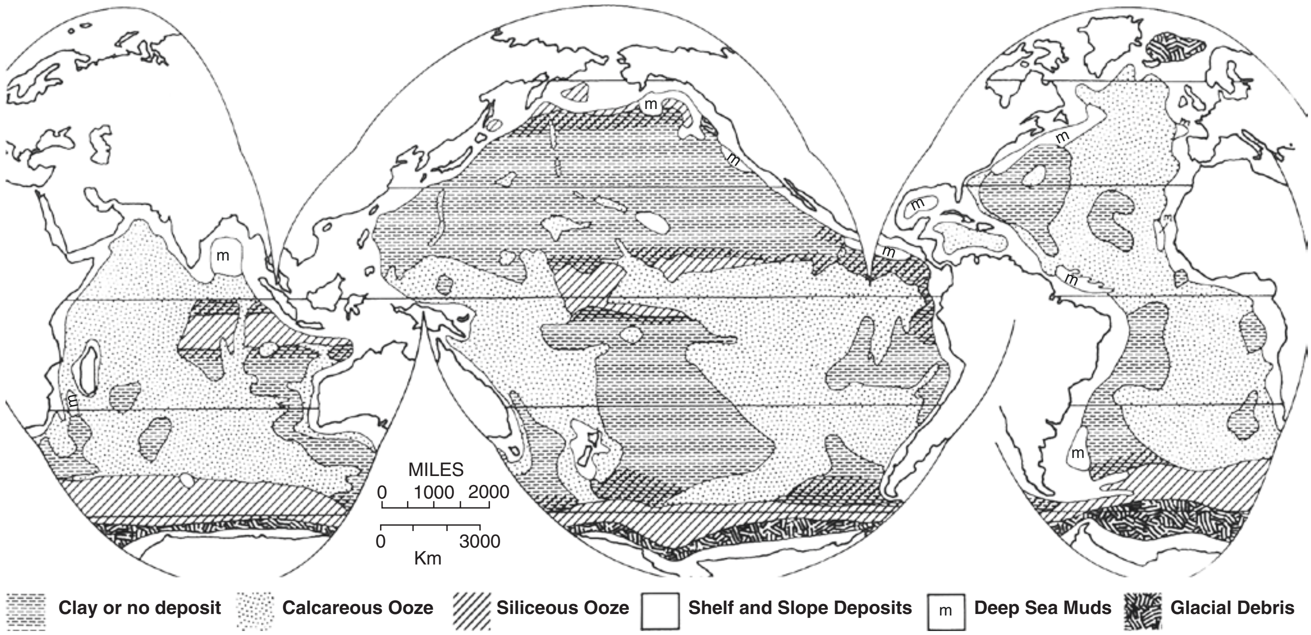

Fig. 10.4 Sediment cover on the deep seafloor. Compiled from many authors (Source as for Table 10.1)

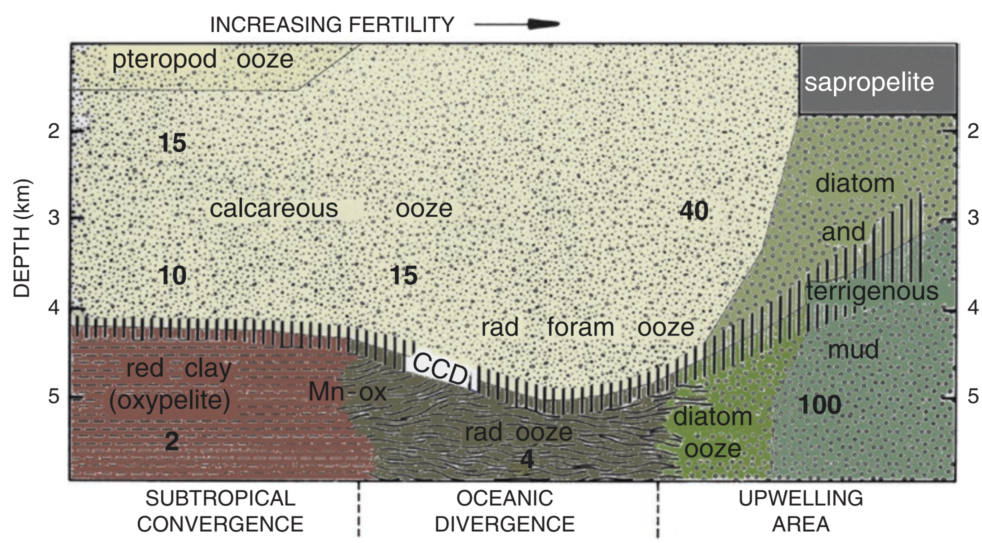

Fig. 10.5 Distribution of major facies in a depth-fertility frame, based on sediment patterns in the eastern central Pacific. Numbers denote typical modern sedimentation rates in mm/millennium (m/million years) (W.H.B., 1974, in C.A. Burk and C.L. Drake (eds.). The *Geology of Continental Margins*. Springer, Heidelberg, Berlin, New York) (Pattern after E. Olausson; modified; color here added)

Chap. 12) that this pattern arose sometime in the Middle Miocene reversing the previously existing conditions. Potential causes for the great shift in sediment patterns ("the silica shift" of the Princeton geologist G. Keller) are elusive but presumably are linked to the overall cooling of the planet changing deep-sea circulation and hence affecting silica deposition in upwelling systems. The new pattern may have arisen in connection with major ice buildup in Antarctica at the time.

The Atlantic-Pacific difference is most obvious in the northern parts of the two basins – the oceanic regions farthest away from each other. The difference in carbonate deposition is truly striking. It was described in the first half of the twentieth century by the Californian geologist and oceanographer R. R. Revelle (1909–1991). The contrast chiefly consists in the fact that carbonate percentages are higher in the North Atlantic than in the Pacific at all similar depths and that the CCD is unusually deep in the North Atlantic. It seems very reasonable to suppose that the phenomenon is linked to *NADW production*, (i.e., the generation of North Atlantic deep water). This production sets up a type of shallow-water-in and deep-water-out circulation in the North Atlantic that has aspects of "anti-estuarine" circulation. As mentioned, anti-estuarine circulation is well appreciated as creating a carbonate trap, while the opposite ("estuarine") pattern favors silica accumulation. In the Atlantic-Pacific

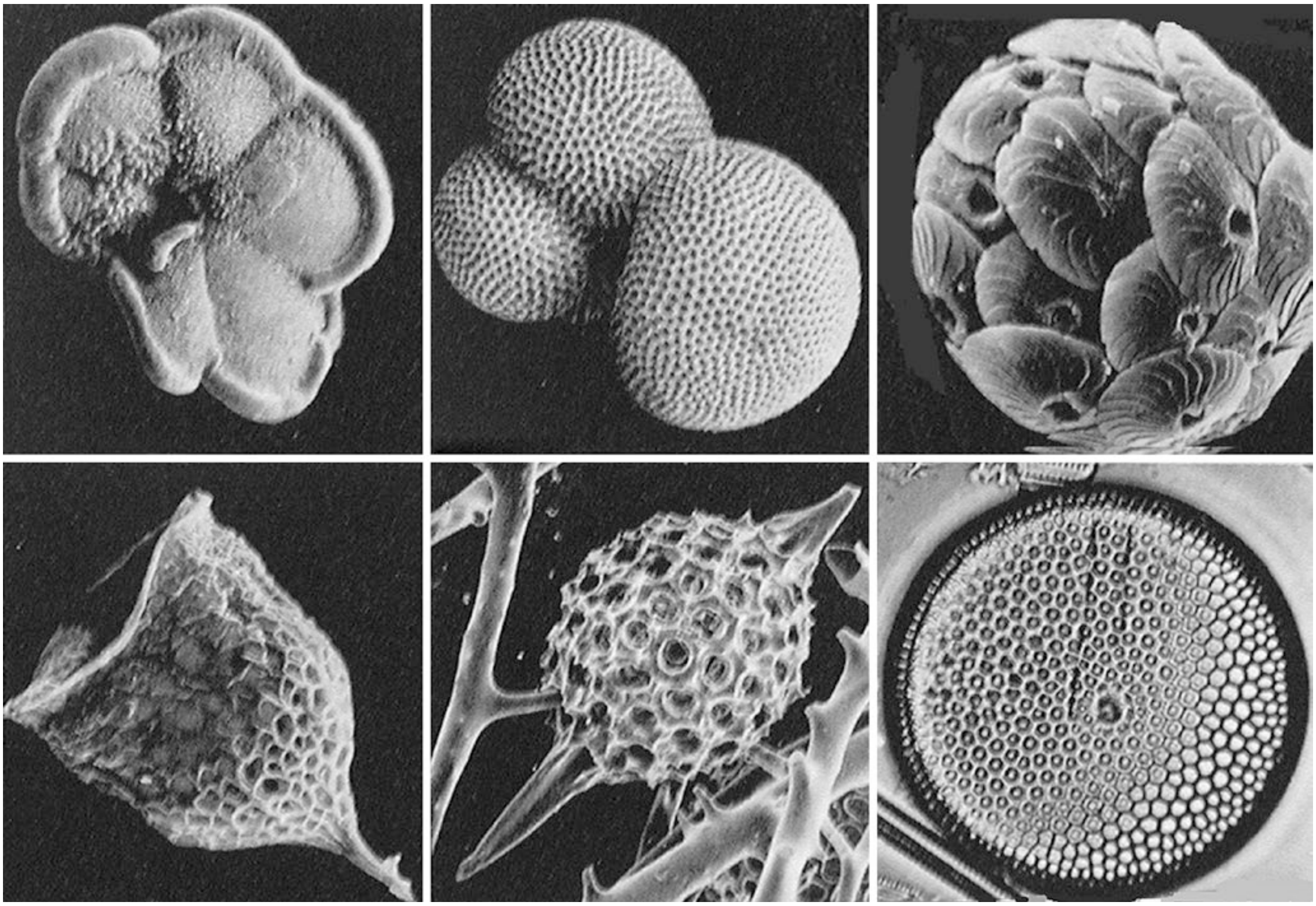

Fig. 10.6 Sediment particles made by shell-bearing plankton. *Upper row*: calcareous forms (two foraminifers, one coccolithophore). On the latter, note the interlocking platelets covering the organism. The platelets are abundant in calcareous ooze. *Lower row*: an organic-walled tintinnid and two siliceous forms (a radiolarian and a cemtric diatom), with diatoms dominant in mud and in coastal upwelling regions (Foraminifers: C. Adelseck, S.I.O.; diatom microphoto courtesy H.J. Schrader, Kiel; others: SEM photos by C. Samtleben and U. Pflaumann, Kiel)

exchange, the northern Atlantic is anti-estuarine, while the northern Pacific is correspondingly estuarine in deep circulation pattern, hence the carbonate-silica dichotomy in sedimentation.

The underlying exchange pattern has been dubbed "basin-basin fractionation." It changes through time, just as is true also for marginal basins. The principle of the peculiar exchange between Atlantic and Pacific basins has found powerful expression in the *conveyor-belt concept* of deep circulation presented in many modern oceanography texts. The conveyor circulation is closely linked to heat transport on the planet. Thus, a question of crucial importance is what happens to the Atlantic-Pacific basin-basin exchange pattern with global warming. A substantial change in the exchange pattern, presumably, will affect the distribution of heat, with several serious consequences. For example, northern Europe could conceivably be cut off from its normal subsidy of heat, now largely delivered by the Gulf Stream system taking warm saline water northward, some of which, upon cooling, returns southward as North Atlantic Deep Water (NADW). The study of sediments with a focus on fluctuations in the heat budget of the deep sea during the ice ages presumably can throw some light on the question of changing exchange patterns. However, one must keep in mind that the ice age sediments of the deep sea typically provide answers for time scales of millennia rather than for those of centuries or decades, owing to bioturbation and a slow sedimentation rate of the archives (ooze).

10.2.4 Sediment Thicknesses and Sedimentation Rates

Given the fact that the seafloor is 60 million years old on average and that typical deep-sea sedimentation rates in the open ocean for oozes and Red Clay range between 1 and 20 mm/millennium, we should expect thicknesses of deep-sea sediments not far from several hundred meters on much

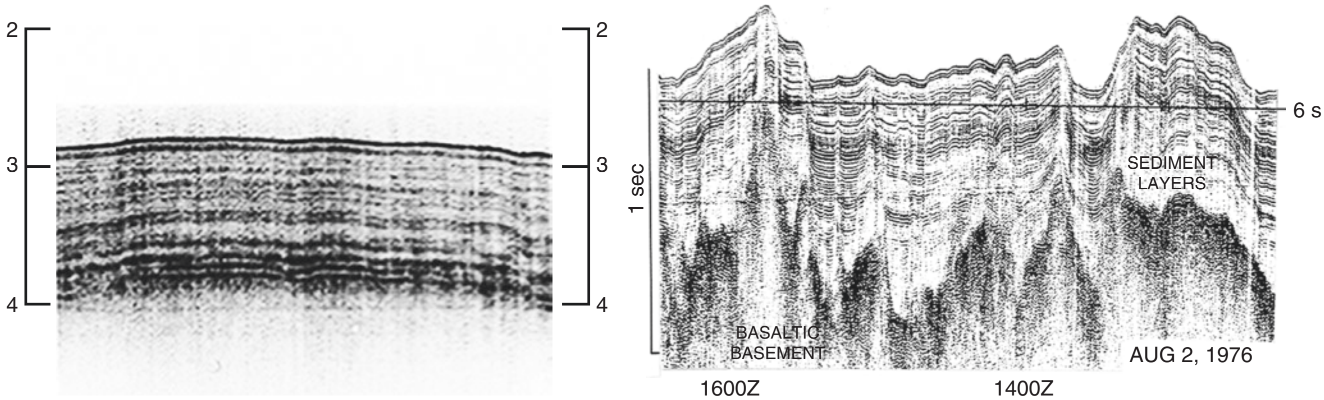

Fig. 10.7 Evidence from seismic echo surveys that deep-sea sediments are surprisingly thin. Thickness scale: echo return time (sound velocity in sediment, 1500–2000 m per s). *Left*: shallow part of Ontong Java Plateau (ca. 100 million years; thickness in excess of 1 km in places); *right*: eastern equatorial Pacific (ca. 50 million years; thickness is ca. 500 m). Sediments show layering (Sources: S.I.O.; left: W. H. B. and T. C. Johnson, 1976. Science 192:785; right: Pleiades Expedition)

of the deep seafloor. In the continental slopes, where rates are ten times higher than on the abyssal seafloor far away from the coastal zone, we should expect thicknesses measured in kilometers. On the whole, what is expected is what was found by drilling into the deep seafloor.

Prior to direct evidence from drilling, the total thickness of sediments on the deep seafloor was known from acoustic methods, that is, subsurface echo sounding or "seismic profiling." It is a method still widely used for mapping sediments. To obtain estimates of sound velocities in ancient subsurface sediments, a more complicated method involving "acoustic refraction" is employed. Given some knowledge of velocities in marine layers, the approximate thickness of a submerged sediment stack can be mapped. In the Pacific rather thick Cenozoic deep-sea sediments (up to 500 m and more) are in the eastern equatorial region, where productivity and hence sediment output are relatively high. Cenozoic and Cretaceous sediments together attain more than 1 km in thickness on Ontong Java Plateau in the western equatorial Pacific. But none of the open ocean thicknesses can compare with the massive deposits off continents in the Atlantic (e.g., off the mouth of the Amazon or the Mississippi). Record thicknesses of more than 10 km, however, are not in the Atlantic but are found in the Indian Ocean, in the Bay of Bengal, which receives debris from the Himalayas.

Generally, then, deep-sea sediment cover is relatively thin, in contrast to the sediment stack of the margins. When discovered, the modest thickness of sediments surprised the geologic community. The oceans were supposed to be a permanent and stable receptacle of continental and volcanogenic debris, with deposits well over a billion years old and with corresponding thicknesses approaching and even exceeding that of the crust. Even in 1959 there were still speculations on the topic. But when doing the appropriate acoustic measurements, the geophysicists Maurice Ewing at Lamont and Russell Raitt (1907–1995) at Scripps, and their collaborators, found that typical sedimentary columns in the deep Atlantic and in the deep Pacific are only a few hundred meters thick, rather than many thousands, a finding subsequently richly confirmed by seismic surveys (Fig. 10.7).

The youth of the seafloor, discovered after the first signs of a thin sediment cover emerged, has resolved the puzzle of thin sediment thickness. Older sediments are missing: they were subducted. Another problem dealing with missing sediment, this one discovered in the 1960s by drilling, has remained unsolved: the problem of missing sections ("hiatuses"). Hiatus development in the deep sea apparently is especially large at times of major change of sedimentation. Compiling early deep-sea drilling results, J. Thiede (Alfred-Wegener Institute, Bremerhaven) and W.U. Ehrmann identified three major hiatus-prone periods in the last 100 million years. Two of them mark changes in facies: about 40 million years ago and near 90 million years ago, toward the end or easing up of a major oxygen crisis in the middle Cretaceous. Also, there is ubiquitous hiatus formation in the earliest Tertiary, right after the end of the Mesozoic (near the "K-T boundary"; see final part of Chap. 13).

To explain hiatuses, geologists commonly invoke pulses of intense erosion during certain periods of climate and circulation change or else large-scale landslides (easily triggered when gas pressure is involved on a continental slope). One thing is known: the overall abundance of hiatuses is correlated with sedimentation rates. The Neogene hiatus formation decreased, as rates of sedimentation increased, thanks to physical weathering from increased ice on land and increased productivity in the sea from a strengthening of winds, desert development on land, and dust supply.

10.2.5 The Pelagic Rain

Sediments on the deep seafloor arrive mainly as a *rain of particles*. The nature of the particle rain has been studied using *sediment traps*, for roughly half of the past century. A crucially important finding is that much of the transport of very small particles is in rapidly sinking fecal pellets, which provides for fast sinking. The sinking of fecal matter can be further enhanced by loading with solid debris of various types (Fig. 7.6). Accelerated sinking by *fecal transport*, and also by aggregate formation, allows even the smallest and the most delicate particles (wind-blown dust, minute coccoliths, small diatom shells) to reach the seafloor (Fig. 7.6). If left to settle individually, such particles would sink at a rate a hundred times slower than observed and at a rate that is inimical to arriving at all, owing to dissolution on the way down.

Despite accelerated sinking, though, the proportion of the primary production that settles from the photic zone (the export production; see Chap. 7) experiences substantial losses during settling. Apparently, many of the pellets disintegrate while sinking or are re-ingested. In any event, proportional amounts that seem missing increase rapidly with depth considered (Fig. 7.7). Settling matter is down to around 1% of production in the open ocean once typical seafloor depths are reached by the rain of particles produced in the sea (Fig. 7.6). Loss of organic matter presumably is accompanied by loss of oxygen in the water, as the organic matter is oxidized. Around continents, within the coastal zone, organic matter supply to the seafloor is enhanced by high production rates, by a short distance to the seafloor, and by the loading of transport agents with mineral matter from erosion. The result of high organic supply is a shortage of oxygen in the seafloor (i.e., black or green sediment, yellow pyrite in ancient rock). The deficiency regarding oxygen interferes with oxidation of the organics, thus enhancing preservation of organic matter.

According to trapping results, seasonal variation in the production of settling matter is of prime importance and is conspicuous in high latitudes (Fig. 10.8). One suspects that interannual fluctuation can be similarly important at times, resulting in pulsed output.

Seasonality of flux to the seafloor implies seasonal feeding by benthic organisms, as well as a pulsed uptake of oxygen. Thus, even though they may live far below and away

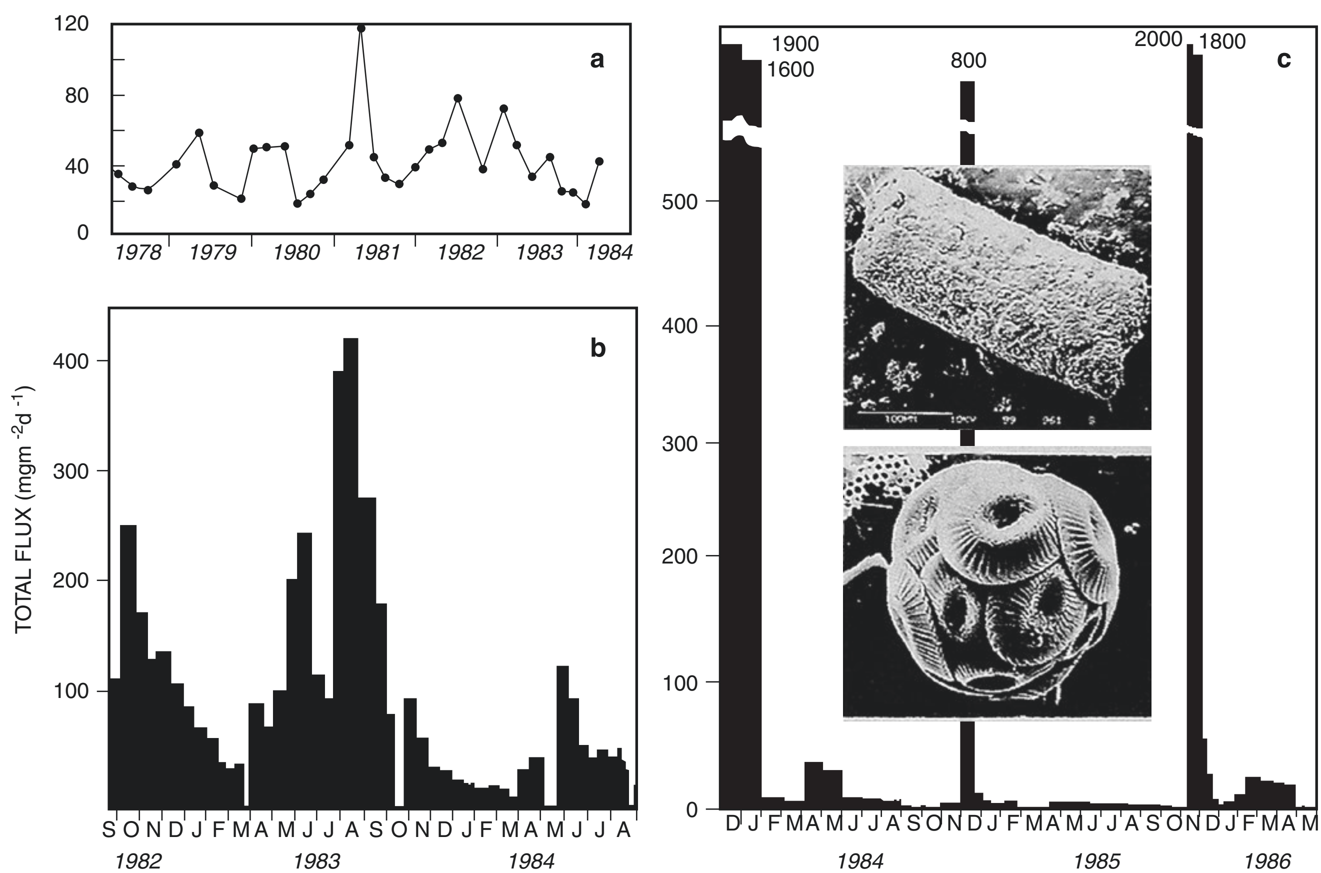

Fig. 10.8 Seasonal and interannual variation in particle flux as observed in traps. (**a**) Sargasso Sea (W.G. Deuser, Woods Hole); (**b**) Gulf of Alaska (S. Honjo, Woods Hole); (**c**) Bransfield Strait, Antarctica (G. Wefer et al., Bremen) (W.H.B. and G. Wefer, 1990. *Global and Planetary Change 3* (3) 245) (Photos: krill fecal string and close-up of coccosphere within the fecal string, courtesy G. Wefer)

from the food source, benthic organisms are subject to feast and famine much like the plankton overhead. Seasonality in the particle rain is especially pronounced in high latitudes, of course, owing to changing supply of sunlight and storm action (nutrient supply). The loading of fecal material with heavy particles, incidentally, is linked to seasons as well, and to fluctuations in climate in general. Seasonal loading implies interesting information on seasonal biological pumping and on apparent oxygen utilization ("AOU") within sediments, especially varved ones (i.e., those with annual layers).

10.3 Calcareous Ooze

10.3.1 General Background

Distribution of the calcareous ooze reflects both production (supply) and chemistry (dissolution) of carbonate particles. Carbonate dissolution increases with depth and also with the supply of organic carbon being oxidized. Organic carbon yields carbonic acid upon oxidation, a compound that attacks carbonate. At the *carbonate compensation depth (CCD)*, the rates of supply and rates of removal of carbonate are balanced: above this boundary there is calcareous ooze; below it we have reddish brown clay (Fig. 10.2).

A most instructive sampling method for the uppermost sediments on the seafloor is box coring, which yields massive samples of calcareous ooze when done in elevated portions of the seafloor, such as Ontong Java Plateau in the western equatorial Pacific (Fig. 10.9). A vertical cut through the material (using a plain metal sheet) creates an exposure that can be washed and studied for disturbance by burrowing and various related benthic activities. On the top of the plateau, where preservation of the shells is excellent and the shear strength of the sediment has not suffered reduction from carbonate dissolution, there are many vertical burrows in the sediment, burrows of a type that are not seen as the CCD is approached at lower elevations. Presumably burrows are modified and destroyed by downslope creep, similar to the creep seen in soils on mountain slopes.

A dark zone appears near the bottom of the profile at the transition between buff-colored modern sediment and older material. The older sediment below the dark zone may have a greenish hue within its color. The dark zone is widespread. It was analyzed in box cores taken in the eastern equatorial Pacific, where it turned out to be rich in iron and manganese. The metals, when in a reduced state, are soluble and thus mobile and move upward with interstitial waters. Both metals precipitate in oxygen-rich conditions, especially the iron. Thus, the dark zone may indicate mobilization of iron and manganese in old organic-rich sediment (presumably glacial in age) and precipitation in (oxygen-rich and organic-poor) modern (postglacial) sediment. The pattern suggests that increased glacial productivity and a drop of supply of organic matter in the transition from glacial time to the Holocene are responsible. Perhaps also the residence time and nature of the bottom water changed, owing to melt water input during deglaciation.

Increased glacial productivity, in any case, is a message from the change in shells and skeletons seen in the sub-cores taken within the boxes and in many other places. It is not difficult to document the changes, because calcareous oozes are largely made of biogenic matter that contains the information. In the easily studied sand fraction, it is shells of planktonic foraminifers. In the hard-to-access clay fraction, it may

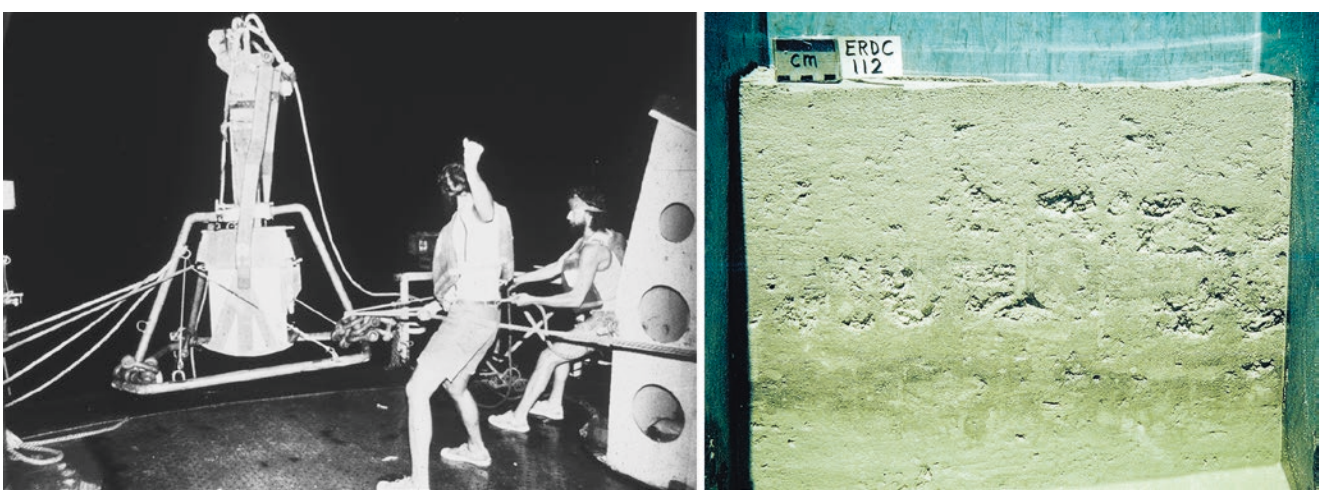

Fig. 10.9 Recovery of calcareous deep-sea sediment by box corer. *Left*: operation of the device, developed from a similar one used by H.E. Reineck (erstwhile director of the Senckenberg Museum in Frankfurt) in the wadden of the North Sea. Lines attached to the equipment are reducing the swinging of the heavy instrument (Photo Tom Walsh, S.I.O.). *Right*: sediment within the box. Note the evidence for a large change of conditions in the lower part of the sediment (Photo W.H.B. and J.S. Killingley, S.I.O.)

be coccoliths (or "nannofossils"), unless we are dealing with carbonate-free "Red Clay." Nannofossils are difficult to identify in a light microscope, but their great abundance holds enormous amounts of environmental information for the expert (in addition to biostratigraphic information useful in dating ancient sediments).

10.3.2 Dissolution of Carbonate Shells

The ultimate fate of calcareous shells settling on the seafloor below the CCD, where supply is compensated by removal, is to dissolve and disappear (Fig. 10.9). The level of disappearance, the *CCD*, can be mapped, once the "critical" carbonate content is defined (the "critical" depth is close to but not equal to the level of zero carbonate; zero would be hard to work with).

Inspection of the CCD map (Fig. 10.10) shows the aforementioned great difference between the Pacific and Atlantic, a significant deepening along the equatorial Pacific and a distinct shallowing in the coastal ocean presumably due to the high supply to the seafloor of organic matter there. The Pacific-Atlantic contrast reflects deep circulation in agreement with the estuarine nature of the northern Pacific and the anti-estuarine one of the northern Atlantic. The deepening of the CCD at the equator indicates increased delivery of carbonate without equally increased delivery of organic matter whose oxidation would destroy carbonate. The organic matter that must come with increased carbonate supply presumably is largely oxidized on the long way down to the seafloor. The removal of organic matter is seen in trapping results and also is implied in the fact that elevated production causes an upward shift of the CCD in the coastal ocean and downward displacement in the deep sea. The main factor, the increasing dissolution with depth, was documented by experiment by the US American geologist and geochemist M.N.A. Peterson (1929–1995) and later chief manager of the Deep Sea Drilling Project. His experiments, and others done with his help, are crucial for understanding the patterns of deep-sea sedimentation documented by John Murray more than a century ago.

What then is the ultimate reason for the presence of the CCD?

It is a matter of geochemical balance. For carbonate (and many other ingredients of deep-sea sediments), the amount available for deposition is fixed by the influx of relevant elements to the ocean from weathering on the continents and from hydrothermal sources. The shell supply to the ocean floor that exceeds the overall influx ultimately has to deplete the sea of calcium carbonate, which results in (pressurized) bottom waters that are sufficiently undersaturated to dissolve the excess supply of calcium carbonate to the seafloor. From this simple bookkeeping concept, it can be readily inferred that, through geologic time, an overall increase in productivity leads to an overall increase in dissolution of carbonate and vice versa. Of course, we must be careful not to extrapolate too far back in time (e.g., beyond the Neogene) when using the present ocean as a model for the past. The geochemical elements of the system change through time, changing the background information accordingly. Before the Neogene (which starts about 24 million years ago), we are dealing with a different planet.

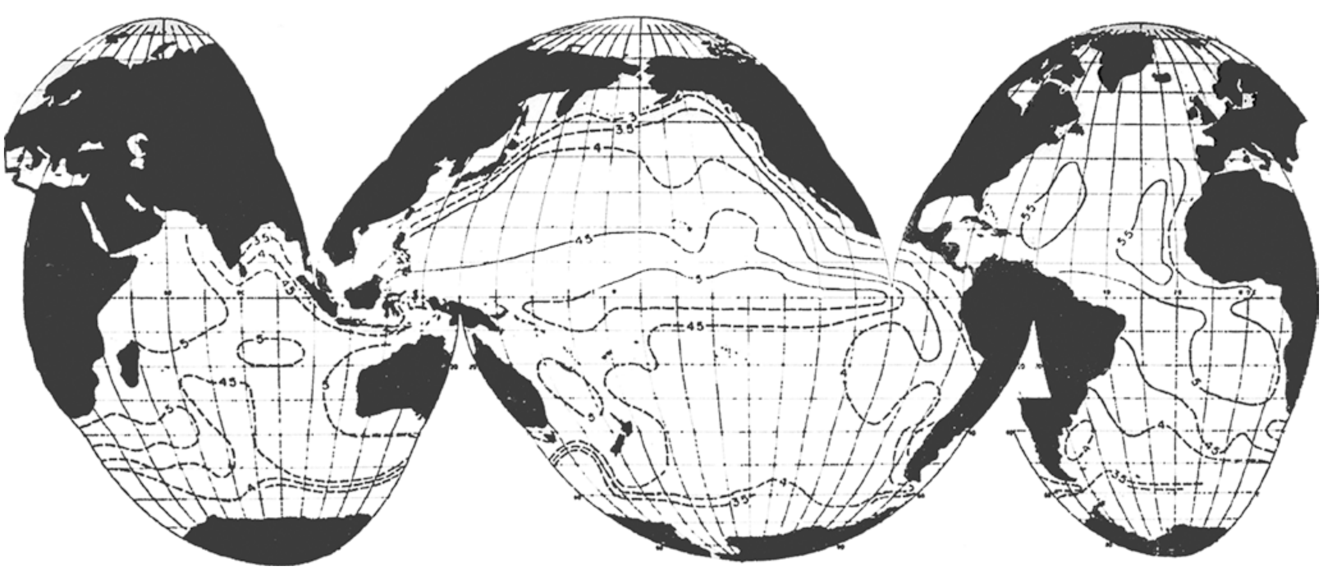

Fig. 10.10 Topography of the carbonate compensation depth (CCD). The seafloor facies boundary (in places extrapolated) between calcareous sediments and sediments with no or very little carbonate. Numbers: depth in km (W.H.B. and E.L. Winterer, 1974. In: K.J. Hsű and H. Jenkyns (eds.) Pelagic Sediments on Land and Under the Sea, Spec. Pub. Int. Assoc. Sedim.1. Blackwell Scientific, Oxford, UK)

Dissolution of carbonate beyond the coastal ocean is largely a matter of depth of deposition, as reflected in the existence of the CCD of the open sea. There is evidence, as surmised by W. Schott and by F. Phleger (S.I.O.), that carbonate dissolution acts differentially toward fossils within calcareous ooze. Increased removal of delicate foraminifer shells at a critical depth has given rise to the concept of *lysocline* (Fig. 10.11, left panel). In essence, the lysocline is much like a CCD for delicate foraminifer shells; hence, the lysocline is well above the CCD. Recording the relative abundance of dissolution-resistant shells in a sediment sample makes it possible to assign a preservation index. When plotting such an index for glacial time and for the Holocene (the last 10,000 years) in the southern Atlantic (Fig. 10.11, right panel), one finds that preservation was poorer than today during the last glacial between 4 and 4.5 km depth within this region. Apparently, both lysocline and CCD stood some 500 m shallower then than during the Holocene. If the associated water-mass boundary in the South Atlantic shifted in similar fashion, the Antarctic bottom water was thicker then by 500 m, and the layer of NADW on top of it was correspondingly reduced. There is supporting evidence from carbon isotopes of benthic foraminifers in sediments of the deep North Atlantic for a reduction of NADW production during the last glacial.

10.3.3 The Carbonate Compensation Depth: Shift in the Pacific

In the tropical Pacific, both preservation signals and carbonate content shifted through vertical ranges that exceeded 500 m in the Pleistocene, but the sense of the change is the reverse from the one found in the South Atlantic. In the Pacific, a relationship to deep-ocean stratification and circulation, if any, is not obvious. To explain the shift, geochemical balance arguments involving the changing availability of shelves presumably have to be invoked. One may assume that lysocline shifts tend to run parallel to those of the CCD; after all, the lysocline is a type of CCD for sensitive foraminifers. That the vertical distance between lysocline and CCD stays entirely unchanged, however, may not be assumed.

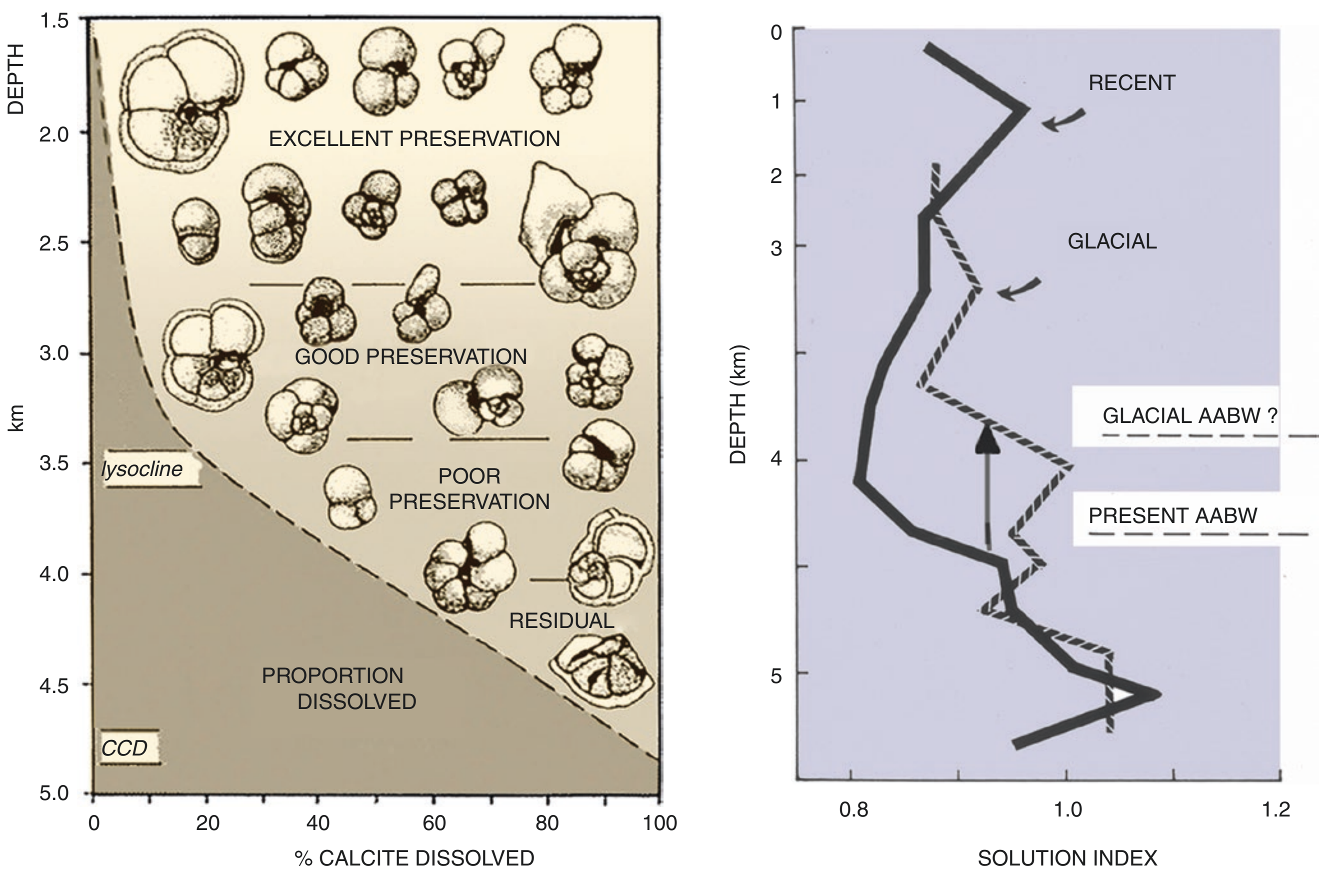

Fig. 10.11 Preservation patterns for planktonic foraminifers on the deep seafloor. *Left*: generalized sketch of distribution of shells in the central Pacific. Drawings courtesy F.L. Parker, S.I.O. (W.H.B., 1985. Episodes 8:163). *Right*: Foraminifer dissolution index in the South Atlantic, in the Holocene and in the last glacial maximum. (W.H.B., 1968. Deep-Sea Research 15:31) *Arrow*: difference in depth between dissolution patterns of the last glacial period and postglacial time

10.3.4 The Global Carbonate Dissolution Experiment

Humans are engaged in a global experiment involving carbonate dissolution. We (in the industrial nations mainly, recently joined by other nations favoring a rapidly expanding economy) are burning enormous amounts of coal and oil at an increasing rate. Large-scale deforestation is proceeding in the tropics and elsewhere for the sake of agricultural development and to obtain wood products and fuel. The carbon dioxide resulting from burning coal, oil, and natural gas, and from destroying forests, enters the atmosphere, from where it is redistributed to other reservoirs, including those in the sea. A doubling of the natural background of carbon dioxide content of the air is projected to occur within the present century, given present trends. Large amounts of the gas are entering the sea. Eventually, according to some estimates, up to ten times the original CO_2 could be added to the atmosphere over the next few centuries, assuming all of the commercially available fossil fuel is burned.

In the long run, reactions at the carbonate-covered seafloor should neutralize most of the industrial carbon dioxide. The formulation of the anticipated reaction on the deep seafloor is relatively simple.

It says that shell carbonate, water, and carbon dioxide react to make (dissolved) ions of calcium and bicarbonate. What is difficult is to construct the timing of the events foreseen; the events per se are readily forecast. For clarity, the events will for sure come; it remains unknown just when.

Unfortunately, the time scale for the process is such that it allows serious damage to the environment during the wait for this (calming) negative feedback to do its work (i.e., carbonate dissolution). Much of the damage done is irreversible on relevant time scales, especially damage involving the unavoidable rise of sea level or extinction. Assessing the rates at which the relevant dissolution reactions are and will be proceeding in the various marine environments is challenging and quite difficult. Mixing of the ocean on a 1000-year scale is implicated, as well as the churning of surficial sediment on the seafloor, by benthic organisms (the time scale of mixing and relevant depths below seafloor are poorly known). Dissolution of carbonate on the seafloor becomes ever slower as the reaction proceeds and removes susceptible shell carbonate from the reactive top layer on the seafloor. We are dealing with several poorly understood feedback mechanisms, including those pertinent to climate change itself, on various time scales. What we strongly suspect is that the processes presumed to be involved in negative feedback typically have long time scales, while positive feedback (albedo change and methane release) has short ones: not a welcome situation if true. At this point it is not known whether short-term and long-term feedback tend to have a different sign, as suspected.

10.4 Siliceous Ooze

10.4.1 Composition and Distribution of Siliceous Ooze

Siliceous oozes are biogenic and they are dominated by diatoms. Diatom oozes and siliceous muds are quite common and widespread in areas of high production (Fig. 10.12, left panel); radiolarian ooze is typical mainly for the deposits below the equatorial upwelling area in the eastern equatorial Pacific (Fig. 10.12, right panel)**.** The fact that diatom production is high around continents leads to the formation of a *silica ring* around each ocean basin. In addition, there are latitude-following *silica belts* resulting from oceanic divergences (notably equatorial upwelling) that are linked to

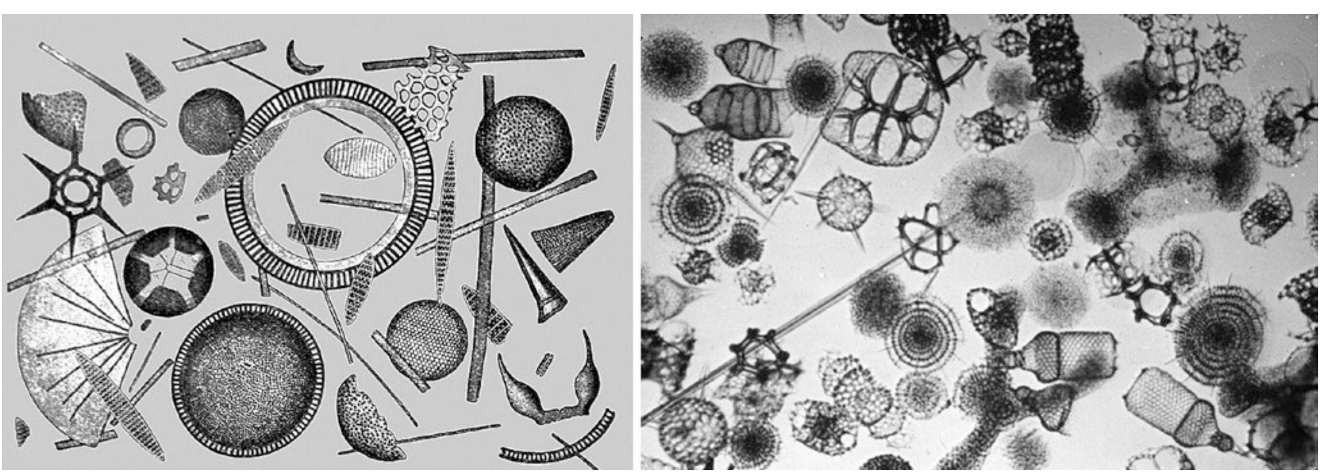

Fig. 10.12 Sand-sized microfossils in siliceous ooze. *Left*: diatom ooze [image from the marine biologist C. Chun via the textbook author O. Krümmel, 1907]; *right*: radiolarian ooze (Microphoto W.H.B., guided by W. Riedel, then S.I.O.)

atmospheric circulation. The regions of divergence have nutrient-rich (hence commonly silicate rich) waters that foster production of microfossil shells, including siliceous ones, such as diatoms and radiolarians. The diluent carbonate commonly is attacked by the organic matter associated with a high supply of diatoms. (In the case of radiolarian ooze, carbonate tends to be removed because of the great depth at which one finds the radiolarian ooze. Above the CCD, on the deep seafloor collecting siliceous fossils, there is siliceous calcareous ooze.)

As outlined in Table 10.1, siliceous oozes bear other constituents as well without losing their main appellation (carbonate, mineral clay, volcanic ash, and others). However, in cases where terrigenous or volcanogenic admixtures are very abundant, we speak of "mud" rather than of "ooze." Siliceous matter being especially abundant in areas of high production (Fig. 7.4) and muds being prominent on the seafloor of the coastal zone, siliceous muds are quite common in the modern ocean off continental margins. Siliceous sediment is not necessarily typical only for the coastal ocean, though, nowhere is there more siliceous sediment than around Antarctica.

10.4.2 Controlling Factors

In analogy to calcareous ooze, the concentration of siliceous fossils in the sediment is a function of (1) the rate of production of siliceous organisms in the overlying waters, (2) the degree of dilution by matter other than biogenic silica (chiefly terrigenous, volcanogenic, and calcareous particles), and (3) the extent of dissolution of the siliceous microfossils, much of which apparently occurs shortly after deposition, that is, within millennia (Fig. 7.15). However, its intensity does not have a simple relationship to depth, as is the case for carbonate for which great depth is a hostile place of deposition.

Siliceous production, on the whole, is a version of the general plankton production discussed in Chap. 7. To obtain an estimate of the amount of silica precipitated in the upper waters, one might multiply the measured amount of organic production with the percentage of solid silica in the organic matter found in suspension in the productive zones (some marine geologists have done this). Presumably the procedure only yields a rather rough estimate, though. We cannot assume that siliceous phytoplankton has rates of growth and reproduction that are identical to the corresponding rates of other plankton. An overall fixation rate of around 200 g SiO_2 per square meter per year has been suggested. A range from less than 100 g (in the central gyres) to more than 500 g (off Antarctica) seems to be a reasonable guess. Of a fixation of 200 g/m^2 year, only about one half of 1 % can end up in sediments if river input is taken as the source of silica and if geochemical balance is to be maintained. If we assume a contribution from seawater-basalt reactions equal to that of rivers, the output can be doubled without violating the bookkeeping balance. Thus, an acceptable global estimate for silica accumulating on the seafloor is then near 1% of production. The implication is that the greater part by far of silica production has to be redissolved either in the water column or on the seafloor, or within the sediment. Accumulation represents a smallish (and therefore highly selective) portion of what is produced.

According to the late geochemist John Martin and associates in Monterey, California, and other scientists studying the matter, diatom production is stimulated by the supply of trace amounts of iron (largely by recycled iron at the margins issued by oxygen-poor sediments and by rivers and dust storms coming from continents). Dilution provides for mixtures, resulting in siliceous mud near continents and in areas of volcanic activity. Dissolution of siliceous material, in the present ocean, seems to be especially vigorous in shallow waters. The evidence for dissolution on the seafloor is striking. Some shells of rather common diatoms, abundantly produced in the sunlit zone, are hard to find on the seafloor. Many siliceous shells and skeletons show signs of poor preservation. In general, silicoflagellates and diatoms tend to dissolve well before robust radiolarians do. Certain sponge spicules seem to be especially resistant. The range of susceptibility to dissolution in microfossils is large enough it apparently can interfere with an assessment of when, in geologic time, silicoflagellates and diatoms first appeared on the planet.

Because of the overriding importance of Antarctic opal deposition (mainly in the shape of diatom debris), the preservation of siliceous shells in the rest of the ocean must to a large degree depend on how much the Antarctic ocean is able to extract from the water column for deposition on its own surrounding seafloor and how much of that is being recycled to diatom-producing upwelling areas. In any case, the modern ocean seems to be rather sensitive in this regard. The evidence consists in strange ice-age patterns, with increased global production resulting in a decreased supply of siliceous shells to the seafloor off Namibia (*Walvis Paradox*; see the next chapter).

10.4.3 Acoustics and Silica Geochemistry of Cenozoic Sediments at the Ontong Java Plateau

The dissolution of opaline shells and skeletons within the surficial sediment layer delivers silica to bottom waters (see Fig. 7.15). From the fact that older bottom waters have the higher concentrations of silicate, we can draw an obvious conclusion: The reason that silica contents in deep water are relatively low cannot be the general uptake, if any, of dissolved silicate by clay minerals on the seafloor. If this were the case, older water should have less silica than younger water. Instead, presumably, the deep water remembers its depleted condition at the surface (owing to diatoms extracting silica to make shells), and it is on the way to saturation with silica by dissolving diatom shells on its travels that end in the northern North Pacific, where concentrations are highest in the sea. Oxygen content decreases as silicate content of bottom waters increases. (In a warm ocean with a different circulation structure, we may have to consider different processes.)

While most of the within-sediment dissolution apparently occurs in the most recent sediment, considerable buildup of concentrations of dissolved silica is seen in ancient sediments on Ontong Java Plateau in the western equatorial Pacific as well, right down to the bottom of the Oligocene (Fig. 10.13; quantitative time scales are still very approximate, especially in the Oligocene). A drastic drop within the latest Eocene and earliest Oligocene may indicate the precipitation of silicate minerals from interstitial waters, presumably large chert (i.e., microcrystalline quartz; perhaps seeking the company of other chert, still abundant in the middle to late Eocene). At that point, also, there is an important *acoustic boundary*, likely denoting a significant change of sound velocity and density.

The sediments down to the middle to the late Eocene reflectors ("Ontong Java Series") consist largely of calcareous ooze and chalk. On drilling down into the sequence, chert beds first appear near the top of the Ontong Java Series. They are middle to late Eocene in age. The content of dissolved silicate in interstitial waters drops at that horizon, suggesting a marked increase in precipitation at that horizon.

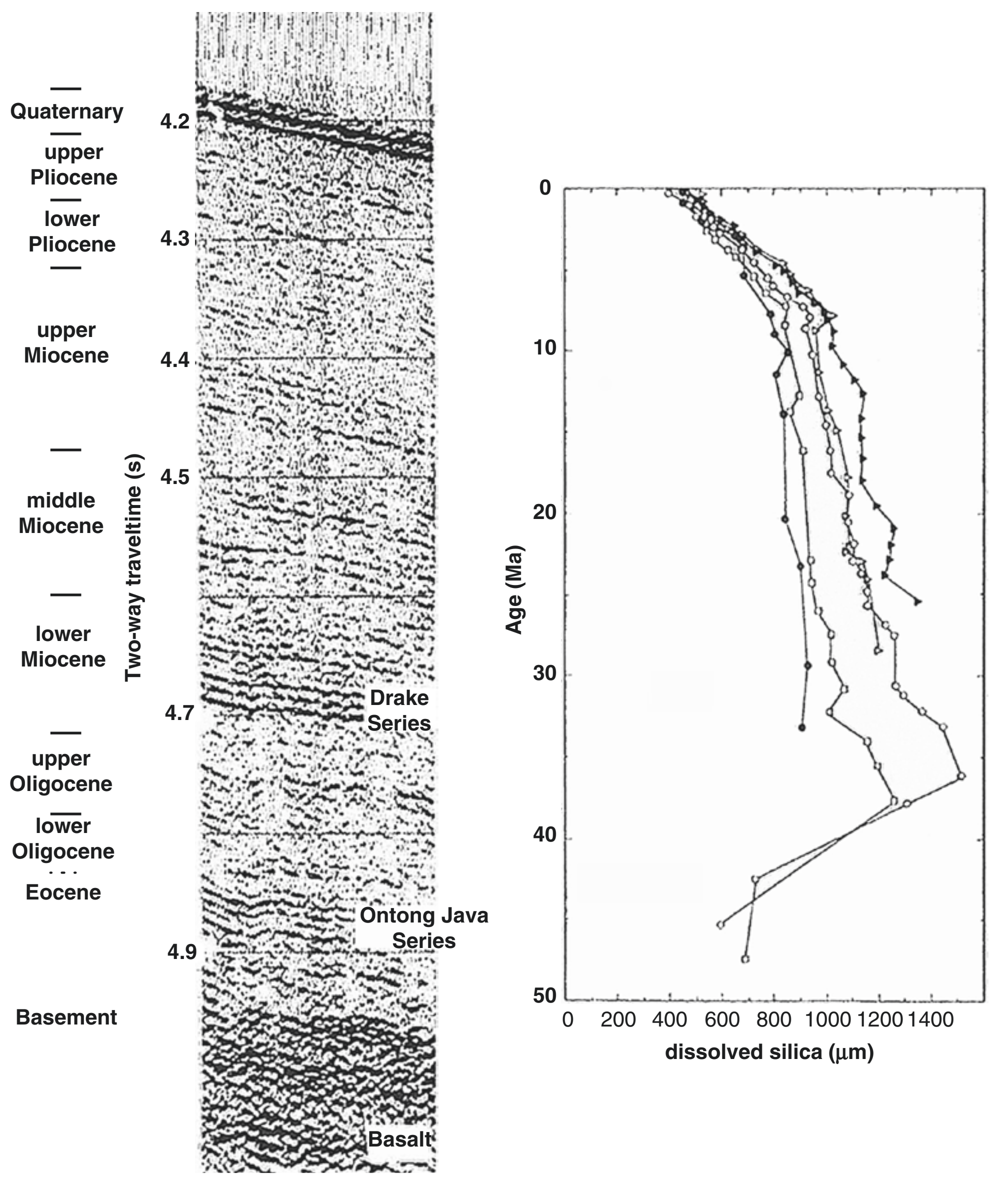

Fig. 10.13 Physical and chemical properties related to silica deposition at ODP Sites on Ontong Java Plateau, western equatorial Pacific. *Left*: acoustic reflectors at Site 805 (vertical scale: two-way travel time of sound, in seconds). *Right*: interstitial water content in Site 805 and several nearby Sites; approximate age in million years; dissolved silica in micro-moles (ODP Leg 130; silicate analyses by M.L. Delaney, UC Santa Cruz; chart of reflectors and microfossil data: shipboard data)

10.4.4 On the Formation of Deep-Sea Chert

The discovery of *chert* (microcrystalline quartz-rich rocks originating from opal) in deep-sea sediments has fascinated geologists and has resulted in much discussion, therefore. Also, it has frustrated them in efforts to recover complete sections of ancient sediments. (Drilling through chert posed some problems for recovery. The effort has been compared with trying to drill through a stack of porcelain dinner plates). The formation of deep-sea cherts appears to proceed from mobilization and re-precipitation of opal. Recrystallization may proceed at various rates depending on the original sediment undergoing alteration. The appearance of chert has been ascribed to both increased volcanism (supply of volcanic ash) and to an increase in diatom production (supply of siliceous microfossils), by different authors working in different locations. Statements on the origin of chert commonly allow for much guesswork, any fossils having largely disappeared as the rock under discussion is recrystallized.

In the western equatorial Pacific and in many other areas of the global ocean, massive chert beds first appear in upper Eocene sediments (i.e., after the Oligocene is penetrated) when drilling down into the seafloor (Fig. 10.12, "Ontong Java Series"). At that level of interbedded limestones and chert, where density of the material changes suddenly and with it the *acoustic impedance* (product of sediment density and sound velocity) sound waves are strongly reflected. A large change in the silica content of interstitial waters indicates loss of dissolved matter to precipitation, presumably to formation of chert and siliceous cement. The measured profiles of dissolved silicate suggest substantial reflux of silica to bottom waters above the seafloor from sediments as old as 10 million years (i.e., from late Neogene sediments). The ongoing loss of silica to bottom waters, of course, is unfavorable for the formation of chert.

The question of why deep-sea chert deposits are concentrated in some geologic periods and not in others is difficult or impossible to answer at this time. Presumably, there is no silver bullet answer, and instead answers are hidden in various elements of the sedimentary silica cycle. Certain ancient chert layers exposed at the margins of the eastern Pacific as radiolarites and ribbon cherts in ophiolites and in mélanges created in subduction zones may well be of turbidite origin, that is, deposited by bottom-hugging sediment-laden downhill flows.

10.5 "Red Clay" and "Clay Minerals"

10.5.1 Early Thoughts on the Origin of "Red Clay"

Of all types of marine sediments, "Red Clay" is uniquely restricted to the deep-sea environment. The bulk of the components consists of extremely fine-grained particles that are difficult to identify, except by highly technical means such as X-ray analysis, a method introduced in the first half of the twentieth century, in decades following discoveries by the physicist Wilhelm Röntgen (1845–1923) in Munich. His X-rays (or "Röntgen" Rays) are familiar from use in the medical sciences and in zoology. They are routinely employed to identify minerals, as well. Murray and Renard, who did not have the benefit of such tools, studied the composition of coarse silt and fine sand particles in the Red Clay, assuming that the results might provide information about the origin of Red Clay as a whole. They found minerals that precipitated on the seafloor, volcanogenic debris in various states of alteration, minute ferromanganese concretions, and traces of biogenic particles such as fish teeth, arenaceous foraminifers, and (in some cases) sponge spicules and radiolarians. In other words, the relatively modest portion of accessible sediment in "Red Clay" was rich in non-calcareous coarse-silt-sized matter of marine origin.

Their finding that the coarser particles in "Red Clay" were non-calcareous confirmed an early suspicion that "Red Clay" is simply what is left over from calcareous ooze, after dissolving the carbonate. The composition of coarse silt and fine sand does not, however, correctly reflect the composition of the clay-sized material. While the decomposition of volcanic material is indeed important in supplying some of the dominant clay minerals ("montmorillonite" or "smectite" and its diagenetic products such as "illite"), there is considerable contribution from continental erosion. The addition of desert dust to the seafloor can be readily inferred from satellite images taken off Africa, with dust being carried all the way to the Caribbean and to the Amazon basin. The transportation of desert dust in winds off Africa was well known to early pioneers of environmental sciences: Charles Darwin wrote about it in his report about the sea voyage on the *Beagle* (1831–1836). Modern voyagers in the area likewise get to know the fine-grained brown dust covering their vessel on occasion (see Fig. 4.3). The naturalist and diatom expert Christian Gottfried Ehrenberg (1795–1876) found silica from grass ("phytoliths") and from freshwater diatoms ("frustules") in dust samples sent to him by Charles Darwin. The occurrence of terrestrial material on certain shallow parts of the deep seafloor at the time strengthened a concept of "Atlantis" (Plato's sunken land) in some (non-geological) minds.

10.5.2 X-Ray Composition of the Clay Fraction

To find out what the "Red Clay" is made of in the clay fraction (how much is of oceanic, how much of continental origin, and how it got to the deep seafloor), one needs to study the clay minerals in the sediment and their distribution and sedimentation rate. Analysis by X-ray diffraction began in the 1930s (by R.R. Revelle in the Pacific and by C.W. Correns in the Atlantic). The method has been systematically applied to deep-sea deposits since. (For a list of dominant clay minerals, see Fig. 10.14 and the Appendix.)

In the North Pacific, one finds a surprising amount of quartz (including on the islands of Hawaii, which are made of volcanic rock) suggesting eolian input from upwind continental deserts, according to S.I.O. geochemist Robert Rex. The *clay minerals* make up the bulk of the clay fraction (ca. two thirds of it, with a median diameter of one thousandth of a millimeter, i.e., 1 μm). Distributional abundance patterns on the seafloor hold clues to origins (Fig. 10.13) (also see Appendix A4).

The patterns suggest that smectite has important sources in oceanic volcanism, at least in the Pacific. The common occurrence of illite in the Atlantic, and especially in continental slopes there, suggests derivation from continental sources. Some of the illite apparently results from diagenesis of smectite within old deep-sea sediment, however. The remaining two important clay minerals, chlorite and kaolinite, seem to be continent derived, one from cold regions (physical weathering) and the other from warm and wet regions (chemical weathering). In general it appears that clay minerals in the "Red Clay" have a surprisingly strong component of continental sources, especially in the Atlantic. In the Pacific, oceanic sources and the "Ring of Fire" presumably supply a rich selection of volcanic materials that decay to smectite (montmorillonite), the dominant clay mineral.

10.6 Hemipelagic Mud

Hemipelagic muds are quite as abundant as the oozes and clays (perhaps more so because they are thicker although not as widespread as the deep-sea facies). Muds are thick owing to their high sedimentation rate, which is some ten times higher than that of calcareous ooze, as mentioned. The composition of the muds is quite different from that of the oozes and the "Red Clay," the muds having a strong admixture of relatively coarse continental weathering products or volcanogenic products and of organic matter. The content of the remains of benthic foraminifers and other benthic organisms accumulates much faster in the mud of oceanic margins than in deep-sea sediments (Fig. 4.11). Much of the mud in continental slopes consists of mineral grains and is brought there by turbidity currents. Shaping is by *contour currents*, that is, currents along the continental margin that stay roughly at the same elevation.

Hemipelagic muds, because of mountain building and general cooling leading to the onset of ice ages in the late Cenozoic, are especially thick in the Neogene (i.e., in the Miocene and later). What we see on seismic profiles on the continental margin is largely of Neogene age, as documented by drilling (Fig. 3.6). The great increase in productivity during post-Eocene time delivered considerable siliceous plankton, much of it in the Middle Miocene (see Chap. 12). Also, substantial amounts of organic matter can be present (typically one to several percent of the sediment). In the Neogene sequences, scientists drill mud largely because high sedi-

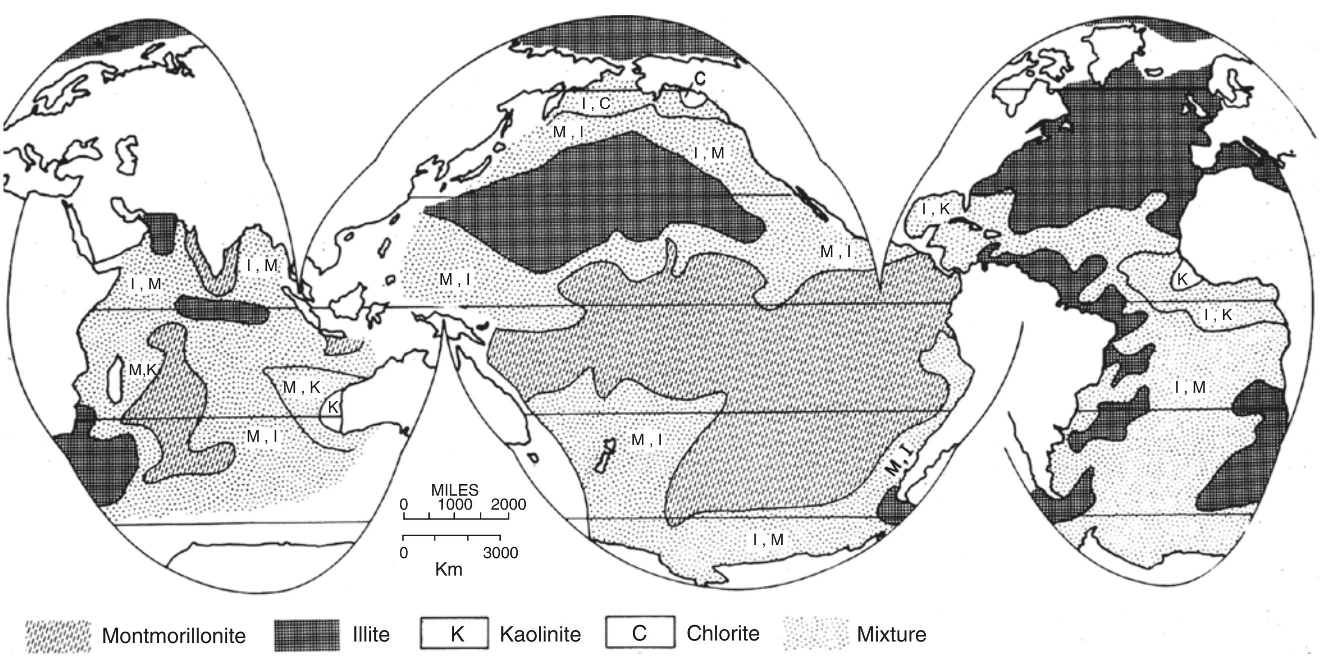

Fig. 10.14 Distribution of the dominant clay minerals in "Red Clay" on the deep seafloor. Compilation from data in works of P.E. Biscaye, J.J. Griffin et al., E.D. Goldberg and J.J. Griffin, D. Carrol, and H.I. Windom (W.H.B., 1974. In: CA. Burk and C.L. Drake (eds.). *The Geology of Continental Margins*. Springer, Heidelberg and Berlin)

mentation rates in the sections promise a detailed geologic history for the late Tertiary. Also, the high productivity displayed in the contents of many muds is attractive for the study of the development of organismic diversity and of evolution in general during that time.

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