1 The Solid Phase of Marine Sediments

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1.1 Introduction

The oceans of the world represent a natural depository for the dissolved and particulate products of continental weathering. After its input, the dissolved material consolidates by means of biological and geochemical processes and is deposited on the ocean floor along with the particulate matter from weathered rock. The ocean floor deposits therefore embody the history of the continents, the oceans and their pertaining water masses. They therefore provide the key for understanding Earth's history, especially valuable for the reconstruction of past environmental conditions of continents and oceans. In particular, the qualitative and quantitative composition of the sedimentary components reflect the conditions of their own formation. This situation may be more or less clear depending on preservation of primary sediment composition, but the processes of early diagenesis do alter the original sediment composition, and hence they alter or even wipe out the primary environmental signal. Hence, only an entire understanding of nature and sequence of processes in the course of sediment formation and its diagenetic alteration will enable us to infer the initial environmental signal from the altered composition of the sediments.

Looking at the sea-floor sediments from a geochemical point of view, the function of particles, or rather the sediment body as a whole, i.e. the solid phase, can be quite differently conceived and will vary with the perspective of the investigator. The "classical" approach – simply applying studies conducted on the continents to the oceans – usually commences with a geological-sedimentological investigation, whereafter the mineral composition is recorded in detail. Both methods lead to a more or less overall geochemical description of the entire system. Another, more modern approach conceives the ocean sediments as part of a global system in which the sediments themselves represent a variable component between original rock source and deposition. In such a rather process-related and globalized concept of the ocean as a system, sediments attain special importance. First, they constitute the environment, a solid framework for the geochemical reactions during early diagenesis that occur in the pore space between the particles in the water-sediment boundary layer. Next to the aqueous phase, however, they are simultaneously starting material and reaction product, and procure, together with the porous interspaces, a more or less passive environment in which reactions take place during sediment formation.

1.2 Sources and Components of Marine Sediments

Ocean sediments are heterogeneous with regard to their composition and also display a considerable degree of geographical variation. Due to the origin and formation of the components various sediment types can be distinguished: Lithogenous sediments which are transported and dispersed into the ocean as detrital particles, either as terrigenous particles – which is most frequently the case – or as volcanogenic particles having only local importance; biogenous sediments which are directly produced by organisms or are formed by accumulation of skeletal fragments; hydrogenous or authigenic sediments which precipitate directly out of solution as new formations, or are formed *de novo* when the particles come into contact with the solution; finally, cosmogenic sediments which are only of secondary importance and will therefore not be considered in the following.

1.2.1 Lithogenous Sediments

The main sources of lithogenous sediments are ultimately continental rocks which have been broken up, crushed and dissolved by means of physical and chemical weathering, exposure to frost and heat, the effects of water and ice, and biological activity. The nature of the parent rock and the prevalent climatic conditions determine the intensity at which weathering takes place. Information about these processes can be stored within the remnant particulate weathered material, the terrigenous detritus, which is transported by various routes to the oceans, such as rivers, glaciers and icebergs, or wind. Volcanic activity also contributes to lithogenous sediment formation, however, to a lesser extent; volcanism is especially effective on the active boundaries of the lithospheric plates, the mid-ocean spreading ridges and the subduction zones.

The major proportion of weathered material is transported from the continents into the oceans by rivers as dissolved or suspension load, i.e. in the form of solid particulate material. Depending on the intensity of turbulent flow suspension load generally consists of particles smaller than 30 microns, finer grained than coarse silt. As the mineral composition depends on the type of parent rock and the weathering conditions of the catchment area, it will accordingly vary with each river system under study. Furthermore, the mineral composition is strongly determined by the grain-size distribution of the suspension load. This can be seen, for example, very clearly in the suspension load transported by the Amazon River (Fig. 1.1) which silt fraction (> 4 - 63 µm) predominantly consists of quartz and feldspars, whereas mica, kaolinite, and smectite predominate in the clay fraction.

It is not easy to quantify the amount of suspension load and traction load annually discharged by rivers into the oceans on a worldwide scale. In a conservative estimative approach which included 20 of the probably largest rivers, Milliman and Meade (1983) extrapolated this amount to comprise approximately $13 \cdot 10^9$ tons. Recent estimations (Milliman and Syvitski 1992) which included smaller rivers flowing directly into the ocean hold that an annual discharge of approximately $20 \cdot 10^9$ tons might even exist.

Under the certainly not very realistic assumption of an even distribution over a surface area of $362 \cdot 10^6$ km^2 which covers the global ocean floor, this amount is equivalent to an accumulation rate of 55.2 tons km^{-2} y^{-1}, or the deposition of an approximately 35 mm-thick sediment layer every 1000 years.

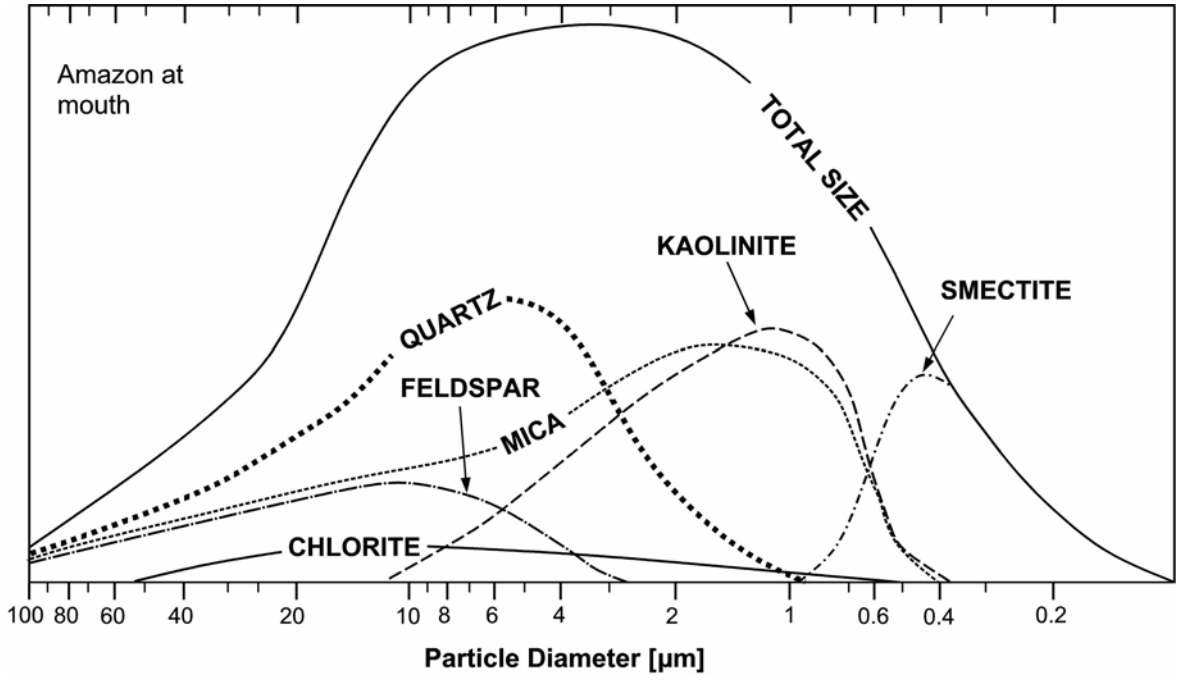


Fig. 1.1 Grain-size distribution of mineral phases transported by the Amazon River (after Gibbs 1977).

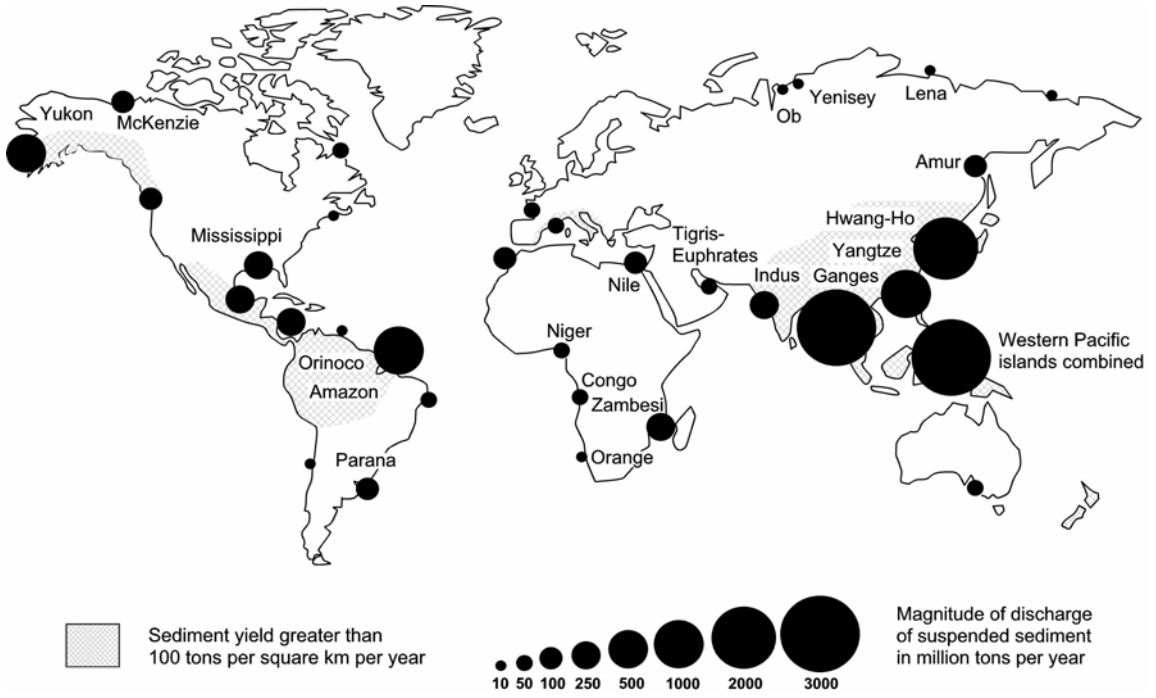


Fig. 1.2 Magnitude of annual particulate sediment discharge of the world's major rivers. The huge amount of sediment discharge in southeast Asia and the western Pacific islands is due to high relief, catchment, precipitation and human activity (Hillier 1995).

Most of the sediment transported to the coastline by the rivers today is deposited on protected coastal zones, in large estuaries, and on the shelves; only a rather small proportion of the sediment is transported beyond the shelf edge and reaches the bottom of the deep sea. The geographical distribution of the particulate discharge varies greatly worldwide, depending on the geographical distribution of the respective rivers, amount and concentration of the suspended material. According to Milliman and Syvitski (1992), the amount of suspension load is essentially a function of the surface area and the relief of the catchment region, and only secondarily does it depend on the climate and the water mass of the rivers. Apart from these influences, others like human activity, climate, and geological conditions are the essential factors for river systems in southeast Asia.

The southeast Asian rivers of China, Bangla Desh, India, and Pakistan that drain the high mountain region of the Himalayan, and the rivers of the western Pacific islands (Fig. 1.2), transport just about one half of the global suspension load discharged to the ocean annually. This must naturally also exert an effect on the sedimentation rates in the adjoining oceanic region of the Indo-Pacific.

Sediment transport by icebergs which calve from glaciers and inland ice into the ocean at polar and subpolar latitudes is an important process for the discharge and dispersal of weathered coarse grained terrigenous material over vast distances. Due to the prevailing frost weathering in nival climate regions, the sedimentary material which is entrained by and transported by the ice is hardly altered chemically. Owing to the passive transport via glaciers the particles are hardly rounded and hardly sorted in fractions, instead, they comprise the whole spectrum of possible grain sizes, from meter thick boulders down to the clay-size fraction.

As they drift with the oceanic currents, melting icebergs are able to disperse weathered terrigenous material over the oceans. In the southern hemisphere, icebergs drift from Antarctica north to 40°S. In the Arctic, the iceberg-mediated transport is limited to the Atlantic Ocean; here, icebergs drift southwards to 45°N, which is about the latitude of Newfoundland. Coarse components released in the process of disintegration and melting leave behind "ice rafted detritus" (IRD), or

"drop stones", which represent characteristic signals in the sediments and are of extremely high importance in paleoclimate reconstructions.

In certain regions, the transport and the distribution carried out by sea ice are important processes. This is especially true for the Arctic Ocean where specific processes in the shallow coastal areas of the Eurasian shelf induce the ice, in the course of its formation, to incorporate sediment material from the ocean floor and the water column. The Transpolar Drift distributes the sediment material across the Arctic Ocean all the way to the North Atlantic. Glacio-marine sedimentation covers one-fifth of present day's ocean floor (Lisitzin 1996).

Terrigenous material can be carried from the continents to the oceans in the form of mineral dust over great distances measuring hundreds to up to thousands of kilometers. This is accomplished by eolian transport. Wind, in contrast to ice and water, only carries particles of finer grain size, such as the silt and clay fraction. A grain size of approximately 80 µm is assumed to mark the highest degree of coarseness transportable by wind. Along wind trajectories, coarser grains such as fine sand and particles which as to their sizes are characteristic of continental loess soil (20-50 µm) usually fall out in the coastal areas, whereas the finer grains come to settle much farther away. The relevant sources for *eolian dust transport* are the semi-arid and arid regions, like the Sahel zone and the Sahara desert, the Central Asian deserts and the Chinese loess regions (Pye 1987). According to recent estimations (Prospero 1996), a total rate of approximately $1\text{-}2 \cdot 10^9$ tons y^{-1} dust is introduced into the atmosphere, of which about $0.91 \cdot 10^9$ tons y^{-1} is deposited into the oceans. This amount is, relative to the entire terrigenous amount of weathered material, not very significant; yet, it contributes considerably to sediment formation because the eolian transport of dust concentrates on few specific regions (Fig. 1.3). Dust from the Sahara contributes to sedimentation on the Antilles island Barbados at a rate of 0.6 mm y^{-1}, confirming that it is not at all justified to consider its contribution in building up deep-sea sediments in the tropical and subtropical zones of the North Atlantic as negligible. Similar conditions are to be found in the northwestern Pacific and the Indian Ocean where great amounts of dust are introduced into the ocean from the Central Asian deserts and the Arabic desert.

According to rough estimates made by Lisitzin (1996), about 84 % of terrigenous sediment input into the ocean is effected by fluvial transport,

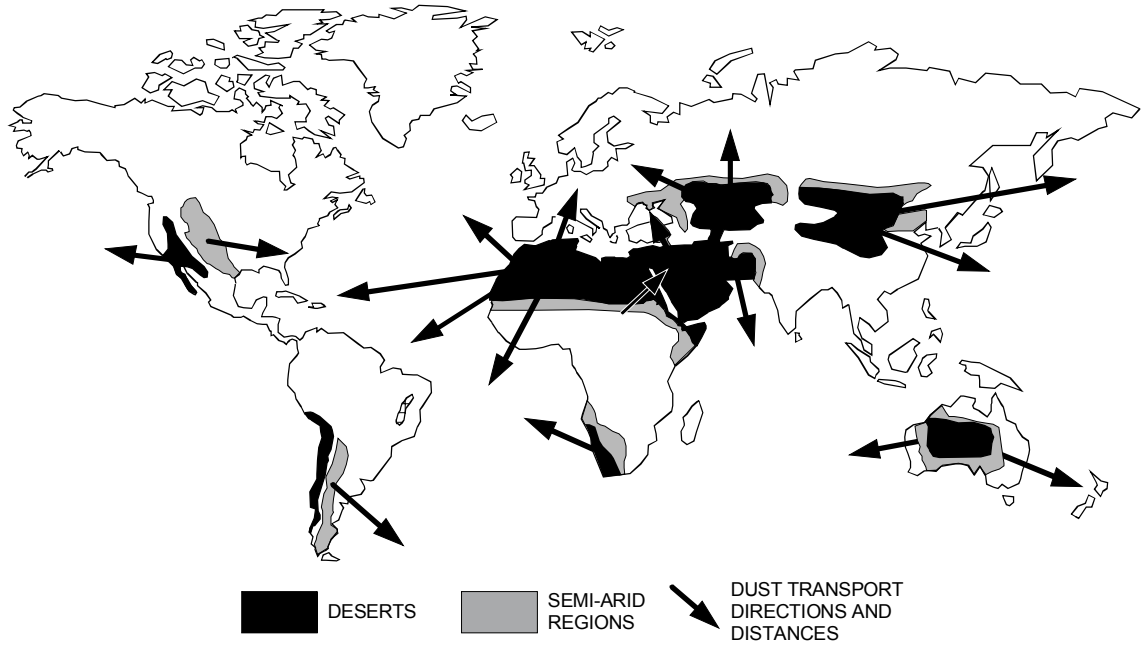


Fig. 1.3 The world's major desert areas and semi-arid regions and potential long-distance eolian dust trajectories and oceanic depocenters (Hillier 1995).

somewhat more than 7 % by eolian transport, and less than 7 % is due to the activity of icebergs.

A distinctly less, but still not insignificant proportion of lithogenous sediment is formed by volcanic activity which is quite often coupled with processes of active subduction at the continental plate boundaries. A large proportion of pyroclastic fragments becomes wind-dispersed over large areas, whereafter they are usually retraced in oceanic sediments as finely distributed volcanic glass. Yet, the formation of single, distinct, cm-thick tephra layers might also occur in the deep sea where they represent genuine isochronous markers which can be used for correlation purposes and the time calibration of stratigraphic units. Layers of ash deposits in the eastern Mediterranean are prominent examples indicative of the eruption of the volcanic island Ischia in prehistoric times of more than 25,000 years ago, and Santorin about 3500 years ago.

Locally, there may be a frequent occurrence of tephra layers and significant concentrations of finely dispersed volcaniclastic material in deep-sea sediments especially in the proximal zones of volcanic activity, like in marginal zones of the modern Pacific Ocean. In a recent evaluation of tephra input into the Pacific Ocean sediments based on DSDP and ODP data Straub and Schmincke (1998) estimate that the minimum proportion of volcanic tephra corresponds to 23 vol.% of the existing Pacific oceanic sediments.

Lithogenous detrital components of marine sediments, despite all regional variability, include only few basic minerals (Table 1.1). With the exception of quartz, complete weathering, particularly the chemical weathering of metamorphic and igneous rock, leads to the formation of clay minerals. Consequently, this group represents, apart from the remaining quartz, the most important mineral constituent in sediments; clay minerals make up nearly 50 % of the entire terrigenous sediment. To a lesser degree, terrigenous detritus contains unweathered minerals, like feldspars. Furthermore, there are mica, non-biogenous calcite, dolomite in low quantities, as well as accessory heavy minerals, for instance, amphibole, pyroxene, apatite, disthene, garnet, rutile, anatase, zirconium, tourmaline, but they altogether seldom comprise more than 1 % of the sediment. Basically, each mineral found in continental rock - apart from their usually extreme low concentrations - may also be found in the oceanic sediments. The percentage in which the various minerals are present in a sediment markedly depends on the grain-size distribution.

The clay minerals are of special importance inasmuch as they not only constitute the largest proportion of fine-grained and non-biogenous sediment, but they also have the special geochemical property of absorbing and easily giving off ions, a property which affords more detailed observation. Clay minerals result foremost from the weathering of primary, rock forming aluminous silicates, like feldspar, hornblende and pyroxene, or even volcanic glass. Kaolinite, chlorite, illite, and smectite which represent the four most important groups of clay minerals are formed partly under very different conditions of weathering. Consequently, the analysis of their qualitative and quantitative distribution will enable us to draw essential conclusions on origin and transport, weathering and hydrolysis, and therefore on climate conditions of the rock's source region (Biscay 1965; Chamley 1989). The extremely fine-granular structure of clay minerals, which is likely to produce an active surface of 30 m^2g^{-1} sediment, as well as their ability to absorb ions internally within the crystal structure, or bind them superficially by means of reversible adsorption, as well as their capacity to temporarily bind larger amounts of water, all these properties are fundamental for us to consider clay sediments as a very active and effectively working "geochemical factory".

Clay minerals constitute a large part of the family of phyllosilicates. Their crystal structure is characterized by alternation of flat, parallel sheets, or layers of extreme thinness. For this reason clay minerals are called layer silicates. Two basic types of layers, or sheets make up any given clay mineral. One type of layer consists of tetrahedral sheets in which one silicon atom is surrounded by four oxygen atoms in tetrahedral configuration. The second type of layer is composed of octahedron sheets in which aluminum or magnesium is surrounded by hydroxyl groups and oxygen in a 6-fold coordinated arrangement (Fig. 1.4). Depending on the clay mineral under study, there is still enough space for other cations possessing a larger ionic radius, like potassium, sodium, calcium, or iron to fit in the gaps between the octahedrons and tetrahedrons. Some clay minerals - the so-called expanding or swelling clays - have a special property which allows them to incorporate hydrated cations into their structure. This process is reversible; the water changes the

Table 1.1 Mineralogy and relative importance of main lithogeneous sediment components.

	relative importance	idealized composition
Quarz	+++	SiO_2
Calcite	+	$CaCO_3$
Dolomite	+	$(Ca,Mg)CO_3$
Feldspars		
Plagioclase	++	$(Na,Ca)[Al(Si,Al)Si_2O_8]$
Orthoclase	++	$K[AlSi_3O_8]$
Muscovite	++	$KAl_2[(AlSi_3)O_{10}](OH)_2$
Clay minerals		
Kaolinite	+++	$Al_2Si_2O_5(OH)_4$
Mica Group		
e.g. Illite	+++	$K_{0.8\text{-}0.9}(Al,Fe,Mg)_2(Si,Al)_4O_{10}(OH)_2$
Chlorite Groupe		
e.g. Chlorite s.s.	+++	$(Mg_{3-y}Al_1Fe_y)Mg_3(Si_{4-x}Al)O_{10}(OH)_8$
Smektite Groupe		
e.g. Montmorillonite	+++	$Na_{0.33}(Al_{1.67}Mg_{0.33})Si_4O_{10}(OH)_2 \cdot nH_2O$
Heavy minerals, e.g.		
Amphiboles		
e.g. Hornblende	+	$Ca_2(Mg,Fe)_4Al[Si_7,Al_{22}](OH)_2$
Pyroxene		
e.g. Augite	+	$(Ca,Na)(Mg,Fe,Al)[(Si,Al)_2O_6]$
Magnetite	-	Fe_3O_4
Ilmenite	-	$FeTiO_3$
Rutile	-	TiO_2
Zircon	-	ZrO_2
Tourmaline	-	$(Na,Ca)(Mg,Fe,Al,Li)_3Al_6(BO_3)_3Si_6O_{18}(OH)_4$
Garnet		
e.g. Grossular	-	$Ca_3Al_2(SiO_4)_3$

volume of the clay particles significantly as it goes into or out of the clay structure. All in all, hydration can vary the volume of a clay particle by 95 %.

Kaolinite is the most important clay mineral of the two-layer group, also referred to as the 7-Ångstrom clay minerals (Fig. 1.4), which consist of interlinked tetrahedron-octahedron units. Illite and the smectites which have the capacity to bind water by swelling belong to the group of three-layered minerals, also referred to as 10-Ångstrom clay minerals. They are made of a combination of two tetrahedrally and one octahedrally coordinated sheets. Four-layer clay minerals, also known as 14-Ångstrom clay minerals, arise whenever a further autonomous octahedral layer emerges between the three-layered assemblies. This group comprises the chlorites and an array of various composites.

Apart from these types of clay minerals, there is a relatively large number of clay minerals that possess a mixed-layered structure made up of a composite of different basic structures. The result is a sheet by sheet chemical mixture on the scale of the crystallite. The most frequent mixed-layer structure consists of a substitution of illite and smectite layers.

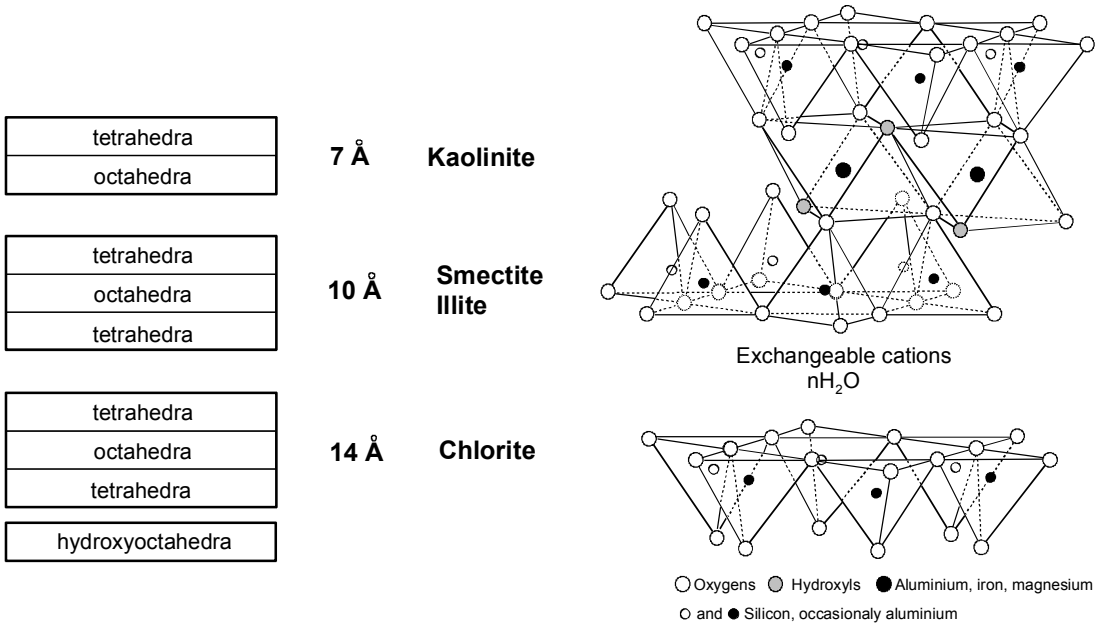


Fig. 1.4 Schematic diagram of clay mineral types: Left: According to the combination of tetrahedral- and octahedral-coordinated sheets; Right: Diagrammatic sketch of the structure of smectite (after Hillier 1995 and Grim 1968).

Kaolinite is a regularly structured di-octahedral two-layer mineral and preferentially develops under warm and humid conditions, by chemical weathering of feldspars in tropical soil. Good drainage is essential to assure the removal of cations released during hydrolysis. Abundance and distribution of kaolinite reflects soil-forming processes in the area of its origin which is optimal in lateritic weathering in the tropics. Owing to the fact that the occurrence of this mineral in ocean sediments is distinctly latitude-dependent, it is often referred to as the "mineral of low latitudes".

Illite is a three-layered mineral of the mica group and not really a specified mineral, instead the term illite refers to a group of mica-like minerals in the clay fraction; as such, it belongs to the most frequently encountered type of clay minerals. Illites are formed as detrital clay minerals by fragmentation in physical weathering. Chemical weathering (soil formation) in which potassium is released from muscovite also leads to the formation of illites. For this reason, illite is often referred to as incomplete mica or hydromica. The distribution of illites clearly reflects its terrestrial and detrital origin which is also corroborated by K/Ar-age determinations made on illites obtained from recent sediments. There are as yet no indications as to *in-situ* formations of illites in marine environments.

Similar to the illites are the cation-rich, expandable, three-layered minerals of the smectite group. The smectites, a product of weathering and pedogenic formation in temperate and sub-arid zones, hold an intermediate position with regard to their global distribution. However, smectites are often considered as indicative of volcanic environments, in fact smectite formation due to low-temperature chemical alteration of volcanic rocks is even a quite typical finding. Similarly, finely dispersed particles of volcanic glass may transform into smectite after a sufficiently long exposure to seawater. Smectites may also arise from muscovite after the release of potassium and its substitution by other cations. There is consequently no distinct pattern of smectite distribution discernible in the oceans.

The generally higher occurrence of smectite concentrations in the southern hemisphere can be explained with the relatively higher input of volcanic detritus. This is especially the case in the Southern Pacific.

Chlorite predominately displays a trioctahedral structure and is composed of a series of three layers resembling mica with an interlayered sheet of brucite (hydroxide interlayer). Chlorite is mainly released from altered magmatic rocks and from metamorphic rocks of the green schist facies as a result of physical weathering. It therefore charac-

teristically depends on the type of parent rock. On account of its iron content, chlorite is prone to chemical weathering. Chlorite distinctly displays a distribution pattern of latitudinal zonation and due to its abundance in polar regions it is considered as the "mineral of high latitudes" (Griffin et al. 1968). The grain size of chlorite minerals is – similar to illite – not limited to the clay fraction (< 4 µm), but in addition encompasses the entire silt fraction (4 - 63 µm) as well.

1.2.2 Biogenous Sediments

Biogenous sediments generally refer to bioclastic sediments, hence sediments which are built of remnants and fragments of shells and tests produced by organisms – calcareous, siliceous or phosphatic particles. In a broader sense, biogenous sediments comprise all solid material formed in the biosphere, i.e. all the hard parts inclusive of the organic substance, the caustobioliths. The organic substance will be treated more comprehensively in Chapter 4, therefore they will not be discussed here.

The amount of carbonate which is deposited in the oceans today is almost exclusively of biogenous origin. The long controversy whether chemical precipitation of lime occurs directly in the shallow waters of the tropical seas, such as the banks of the Bahamas and in the Persian Gulf (Fig. 1.5), during the formation of calcareous ooids and oozes of acicular aragonite, has been settled in preference of the concept of biomineralization (Fabricius 1977). The tiny aragonite

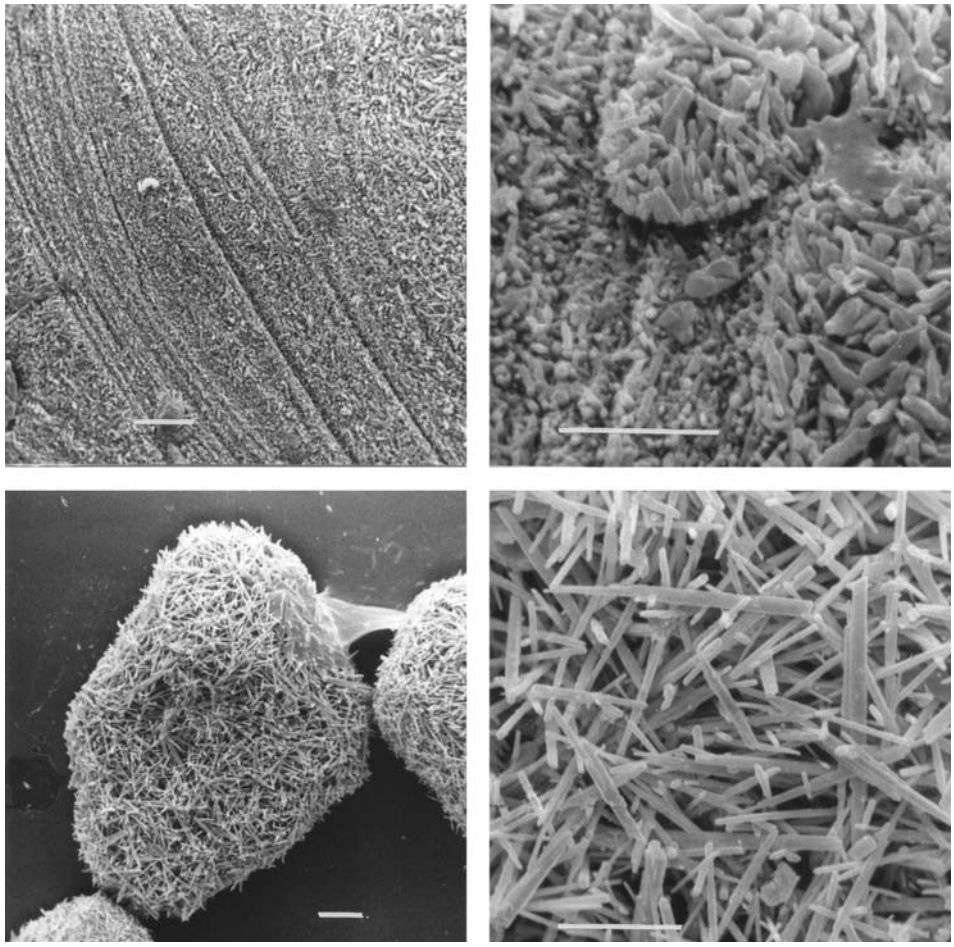


Fig. 1.5 SEM photographs of calcareous sediments composed of aragonite needles which are probably of biogenic origin. *Upper left:* slightly etched section of a calcareous ooid from the Bahamas showing subconcentric laminae of primary ooid coatings; *upper right*: close-up showing ooid laminae formed by small acicular aragonite needles. *Lower left*: silt-sized particle of aragonite mud from the Persian Gulf; *lower right:* close-up showing details of acicular aragonite needles measuring up to 10 µm in length, scalebar 5 µm.

needles (few micrometers) within the more or less concentric layers of calcareous ooids have been considered for a long time as primary precipitates from seawater. However, further investigations including the distribution of stable isotopes distinctly evidenced their biological origin as products of calcification by unicellular algae. Yet, one part of the acicular aragonite ooze might still originate from the mechanical disruption of shells and skeletal elements.

Although marine plants and animals are numerous and diverse, only relatively few groups produce hard parts capable of contributing to the formation of sediments, and only very few groups occur in an abundance relevant for sediment formation (Table 1.2). Relevant for sediment formation are only carbonate minerals in the form of aragonite, Mg-calcite and calcite, as well as biogenic opal in the form of amorphous $SiO_2 \cdot nH_2O$. The sulfates of strontium and barium as well as various compounds of iron, manganese, and aluminum are of secondary importance, yet they are of geochemical interest, e.g. as tracers for the reconstruction of past environmental conditions. For example, the phosphatic particles formed by various organisms, such as teeth, bone, and shells of crustaceans, are major components of phosphorite rocks which permit us to draw conclusions about nutrient cycles in the ocean.

Large amounts of carbonate sediment accumulate on the relatively small surface of the shallow shelf seas – as compared to entire oceans surface – of the tropical and subtropical warm water regions, primarily by few lime-secreting benthic macrofossil groups. Scleritic corals, living in symbiosis with algae, and encrusting red algae consti-

Table 1.2 Major groups of marine organisms contributing to biogenic sediment formation and mineralogy of skeletal hard parts. Foraminifera and diatoms (underlined) are important groups of both plankton and benthos. x = common, (x) = rare (mainly after Flügel 1978, 2004 and Milliman 1974).

	Aragonite	Aragonite + Calcite	Mg-Calcite	Calcite	Calcite + Mg-Calcite	Opal	divers
Plankton							
Pteropods	x						
Radiolarians						x	celestite
Foraminifera	(x)		x	x	x		
Coccolithophores				x			
Dinoflagellates			x				organic
Silicoflagellates						x	
Diatoms						x	
Benthos							
Chlorophyta	x						
Rhodophyta	x		x		(x)		
Phaeophyta	x						
Sponges	x		x			x	celestite
Scleratinian corals	x						
Octocorals	x		x				
Bryozoens	x	x					
Brachiopods				x			phosphate
Gastropods	x	x					
Pelecypods	x	x		x			
Decapods			x				phosphate
Ostracods	(x)		x				
Barnacles	(x)		x				
Annelid worms	x	x	x			(x)	phosphate
Echinoderms			x				phosphate
Ascidians	x						

tute the major proportion of the massive structures of coral reefs. Together with calcifying green algae, foraminifera, and mollusks, these organisms participate in a highly productive ecosystem. Here, coarse-grained calcareous sands and gravel are essentially composed of various bioclasts attributable to the reef structure, of lime-secreting algae, mollusks, echinoderms and large foraminifera.

Fine-grained calcareous mud is produced by green algae and benthic foraminifera as well as by the mechanical abrasion of shells of the macrobenthos. Considerable amounts of sediment are formed by bioerosion, through the action of boring, grazing and browsing, and predating organisms. Not all the details have been elucidated as to which measure the chemical and biological decomposition of the organic matter in biogenic hard materials might lead to the formation of primary skeletal chrystalites on the micrometer scale, and consequently contribute to the fine-grained calcareous mud formation.

It is obvious that the various calcareous-shelled groups, especially of those organisms who secrete aragonite and Mg-calcite, contribute significantly to the sediment formation in the shallow seas, whereas greater deposits of biogenic opal are rather absent in the shallow shelf seas. The isostatically over-deepened shelf region of Antarctica, where locally a significant accumulation of siliceous sponge oozes occurs, however, makes a remarkable exception. The relatively low opal concentration in recent shelf deposits does not result from an eventual dilution with terrigenous material. The reason is rather that recent tropical shallow waters have low silicate concentrations, from which it follows that diatoms and sponges are only capable of forming slightly silicified skeletons that quickly remineralize in markedly silicate-deficient waters.

With an increasing distance from the coastal areas, out toward the open ocean, the relevance of planktonic shells and tests in the formation of sediments increases as well (Fig. 1.6). Planktonic lime-secreting algae and silica-secreting algae, coccolithophorids and diatoms that dwell as primary producers in the photic zone which thickness measures approximately 100 m, as well as the calcareous foraminifers and siliceous radiolarians, and silicoflagellates (Table 1.2), are the producers of the by far most widespread and essential deep-sea sediments: the calcareous and siliceous biogenic oozes. Apart from the groups mentioned, planktonic mollusks, the aragonite-shelled pteropods, and some calcareous cysts forming dinoflagellates, also contribute to a considerable degree to sediment formation.

1.2.3 Hydrogenous Sediments

Hydrogenous sediments may be widely distributed, but as to their recent quantity they are relatively insignificant. They will be briefly mentioned in this context merely for reasons of being complete. According to Elderfield (1976), hydrogenous sediments can be subdivided into “precipitates”, primary inorganic components which have precipitated directly from seawater, like sodium chloride, and “halmyrolysates”, secondary components which are the reaction products of sediment particles with seawater, formed subsequent to *in-situ* weathering, but prior to diagenesis. Of these, manganese nodules give an example. In the scope of this book, these components are not conceived as being part of the “primary” solid phase sediment, but as “secondary” authigenic formations which only emerge in the course of diagenesis, as for instance some clay minerals like glauconite, zeolite, hydroxides of iron and manganese etc. In the subsequent Chapters 11 and 13 some aspects of these new formations will be more thoroughly discussed.

The distinction between detrital and newly formed, authigenic clay minerals is basically difficult to make on account of the small grain size and their amalgamation with quite similar detrital material. Yet it has been ascertained that the by far largest clay mineral proportion – probably more than 90 % – located in recent to subrecent sediments is of detrital origin (Chamley 1989; Hillier 1995). There are essentially three ways for smectites to be formed, which demand specific conditions as they are confined to local areas. Alterations produced in volcanic material is one way, especially by means of hydration of basaltic and volcanic glasses. This process is referred to as *palagonitization*. The probably best studied smectite formation consists in the vents of hydrothermal solutions and their admixture with seawater at the mid-oceanic mountain ranges. Should authigenic clay minerals form merely in recent surface sediments in very small amounts, their frequency during diagenesis (burial diagenesis), will demonstrate a distinct elevation. However, this aspect will not be considered any further beyond this point.

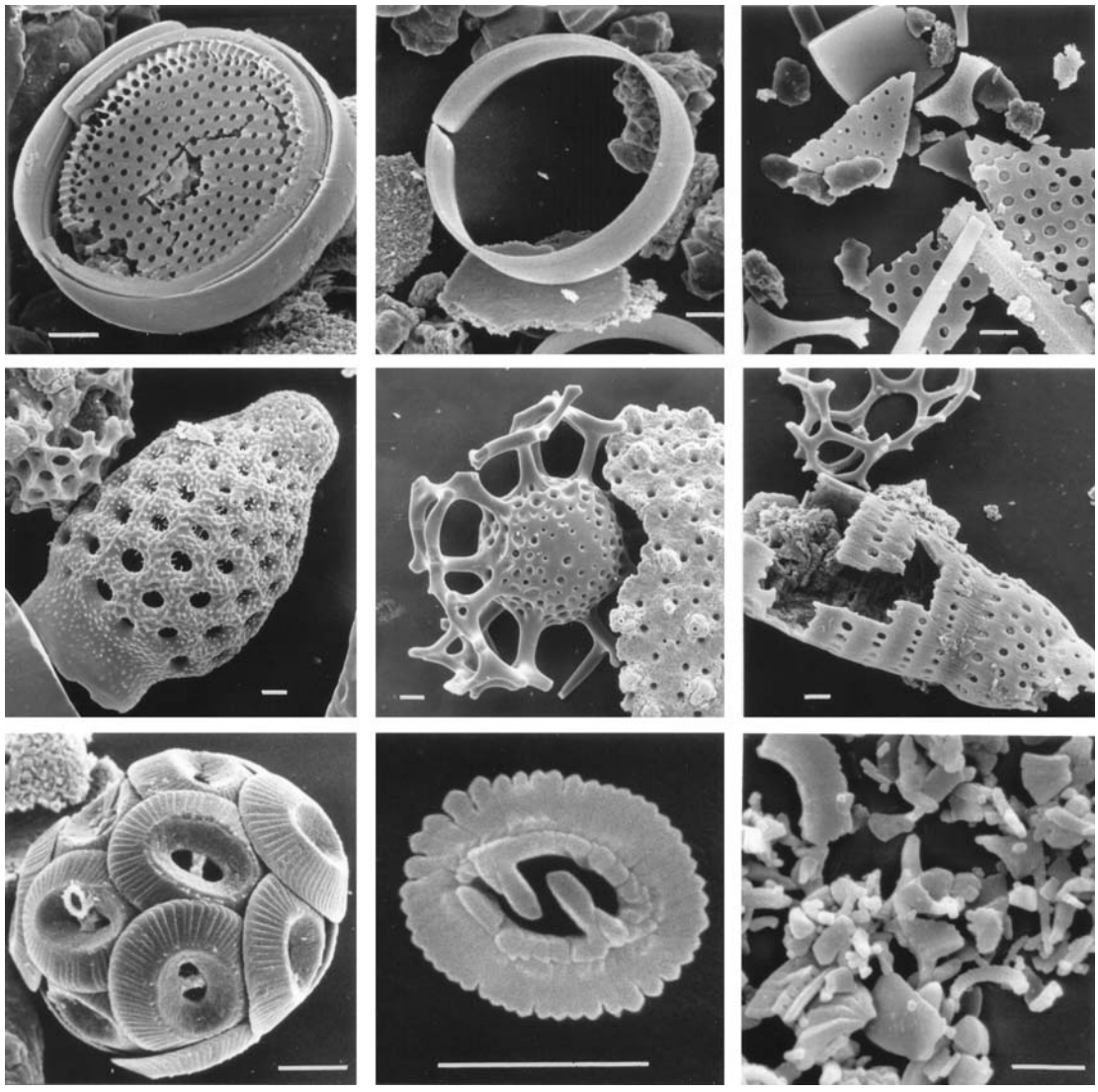


Fig. 1.6 SEM photographs of the important sediment-forming planktonic organisms in various stages of decay. *Above*: Siliceous centered diatoms. *Middle*: Siliceous radiolarian. *Below*: Calcareous coccolithophorid of single placoliths = coccoliths and lutitic abrasion of coccoliths. Identical scale of 5 µm.

1.3 Classification of Marine Sediments

As yet, there is no general classification scheme applicable to marine sediments that combines all the essential characteristics pertaining to a sediment. A large number of different schemes has been proposed in the literature which focus either on origin, grain-size distribution, chemical and mineralogical features of the sediment components, or the facial development of the sediments – all depending on the specific problem under study. The advantages and disadvantages of the various schemes will not be of any concern here. In the following, an attempt will be made to give a comprehensive concise summary, based on the combination of the various concepts.

Murray and Renard (1891) early introduced a basically simple concept which in its essentials differentiated according to the area of sediment deposition as well as to sediment sources. It distinguished (i) “shallow-water deposits from low-water mark to 100 fathoms”, (ii) “terrigenous deposits in deep and shallow water close to land”, i.e. the combined terrigenous deposits from the

deep sea and the shelf seas, and (iii) "pelagic deposits in deep water removed from land". This scheme is very much appropriate to provide the basic framework for a simple classification scheme on the basis of terrigenous sediments, inclusive of the "shallow-water deposits" in the sense of Murray and Renard (1891), and the deep-sea sediments. The latter usually are subdivided into hemipelagic sediments and pelagic sediments.

1.3.1 Terrigenous Sediments

Terrigenous sediments, i.e. clastics consisting of material eroded from the land surface, are not only understood as nearshore shallow-water deposits on the shelf seas, but also comprise the deltaic foreset beds of continental margins, slump deposits at continental slopes produced by gravity transport, and the terrigenous-detrital shelf sediments redistributed into the deep sea by the activity of debris flows and turbidity currents.

The sand, silt, and clay containing shelf sediments primarily consist of terrigenous siliciclastic components transported downstream by rivers; they also contain various amounts of autochthonous biogenic shell material. Depending on the availability of terrigenous discharge, biogenic carbonate sedimentation might predominate in broad shelf regions. A great variety of grain sizes is typical for the sediments on the continental shelf, with very coarse sand or gravel accumulating in high-energy environments and very fine-grained material accumulating in low-energy environments. Coarse material in the terrigenous sediments of the deep sea is restricted to debris flow deposits on and near the continental margins and the proximal depocenters of episodic turbidites.

The development and the hydrodynamic history of terrigenous sediments is described in its essentials by the grain-size distribution and the derived sediment characteristics. Therefore, a classification on the basis of textural features appears suitable to describe the terrigenous sediments. The subdivision of sedimentary particles according the Udden-Wentworth scale encompasses four major categories: gravel (> 2 mm), sand (2 - 0.0625 mm), silt (0.0625 - 0.0039 mm) and clay (< 0.0039 mm), each further divided into a number of subcategories (Table 1.3). Plotting the percentages of these grain sizes in a ternary diagram results in a basically quite simple and clear classification of the terrigenous sediments (Fig. 1.7). Further subdivisions and classifications within the various fields of the ternary diagram are made quite differently and manifold, so that a commonly accepted standard nomenclature has not yet been established. The reason for this lies, to some extent, in the fact that variable amounts of biogenic components may also be present in the sediment, next to the prevalent terrigenous, siliclastic components. Accordingly, sediment classifications are likely to vary and certainly will become confusing as well; this is especially true of mixed sediments. However, a certain degree of standardization has developed owing to the frequent and identical usage of terms descriptive of sediment cores, as employed in the international Ocean Drilling Program (ODP) (Mazullo and Graham 1987).

Although very imprecise, the collective term "mud" is often used in literature to describe the texture of fine-grained, mainly non-biogenic sediments which essentially consist of a mixture of silt and clay. The reason for this is that the differentiation of the grain-size fractions, silt and clay, is not easy to manage and is also very time-consuming with regard to the applied methods. However, it is of high importance for the genetic interpretation of sediments to acquire this information. For example, any sediment mainly consisting of silt can be distinguished, such as a distal turbidite or contourite, from a hemipelagic sediment mainly consisting of clay. The classification of clastic sediments as proposed by Folk (1980) works with this distinction, providing a more precise definition of *mud* as a term (Fig. 1.8).

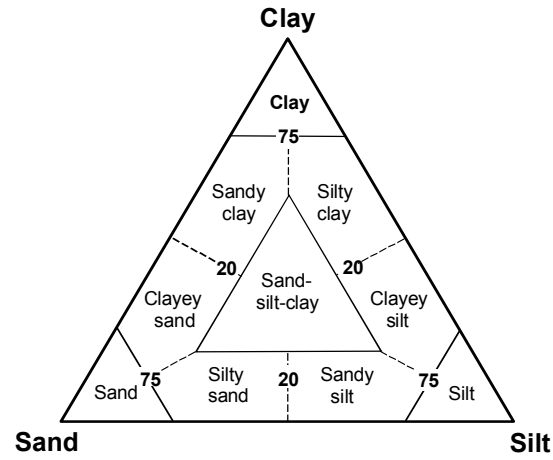


Fig. 1.7 Ternary diagram of sand-silt-clay grain-size distribution showing principal names for siliciclastic, terrigeneous sediments (from Shepard 1954).

mm	Udden-Wenthworth phi values (ϕ)	Terminology
1024	-10	Boulder
512	-9	
256	-8	
128	-7	Cobbles
64	-6	
32	-5	
16	-4	Pebble
8	-3	
4	-2	
3,36	-1,75	
2,83	-1,5	Granule
2,38	-1,25	
2,00	-1,0	
1,68	-0,75	
1,41	-0,5	Very coarse sand
1,19	-0,25	
1,00	0,0	
0,84	0,25	
0,71	0,50	Coarse sand
0,59	0,75	
0,50	1,00	
0,42	1,25	
0,35	1,50	Medium sand
0,3	1,75	
0,25	2,00	
0,21	2,25	
0,177	2,50	Fine sand
0,149	2,75	
0,125	3,00	
0,105	3,25	
0,088	3,50	Very fine sand
0,074	3,75	
0,0625	4,00	
0,053	4,25	
0,044	4,50	Coarse silt
0,037	4,75	
0,031	5,00	
	5,25	
	5,50	Medium silt
	5,75	
0,0156	6,00	
	6,25	
	6,50	Fine silt
	6,75	
0,0078	7,00	
	7,25	
	7,50	Very fine silt
	7,75	
0,0039	8,00	
0,00200	9,0	
0,00098	10,0	**Clay**
0,00049	11,0	

Table 1.3 Grain-size scales and textural classification following the Udden-Wentworth US Standard. The phi values (ϕ) according to Krumbein (1934) and Krumbein and Graybill (1965); $\phi = - \log_2 d/d_0$ where d_0 is the standard grain diameter (i.e. 1 mm).

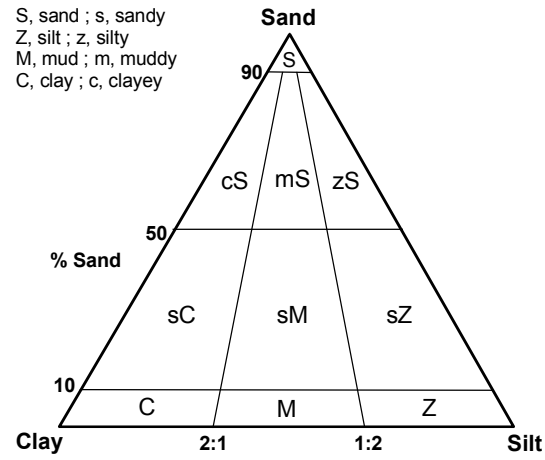


Fig. 1.8 Textural classification of clastic sediments (modified after Folk 1980).

It needs to be stated in this context, that, in order to compare the quantitative reports of published grain sizes, the international literature does not draw the line between silt and clay at 0.0039 mm – as the U.S standard and the Udden-Wentworth scale does (Table 1.3), or the French AFNOR-norm, but sets the limit at 0.002 mm, a value very often found in German literature and complying with the DIN-standards. To add to further diversity, modern publications from Russia mark the silt-clay transition at a grain size of 0.01 mm (e.g. Lisitzin 1996).

1.3.2 Deep-sea Sediments

The sediments in the deep sea consist of only few basic types which in their manifold combinations are suited for the description of a varied facial pattern (Table 1.4). The characteristic pelagic deep-sea sediment far from coastal areas is deep-sea red clay, an extremely fine-grained (median < 1 µm) red-brown clay sediment which covers the oceanic deep-sea basins below the Calcite Compensation Depth (CCD). More than 90 % is composed of clay minerals, other hydrogenous minerals, like zeolite, iron-manganese precipitates and volcanic debris. Such sediment composition demonstrates an authigenic origin. The

small percentage of lithogenic minerals, such as quartz, feldspar and heavy minerals, confirms the existence of terrigenous components which in part should have originated from eolian transport processes. The biogenic oozes represent the most frequent type of deep-sea sediments; they mainly consist of shells and skeletal material from planktonic organisms living in the ocean where they drizzle from higher photic zones down to the ocean floor, like continuous rainfall, once they have died (Fig. 1.6). The fragments of the calcareous-shelled pteropods, foraminifera, and coccolithophorids constitute the calcareous oozes (pteropod ooze, foraminiferal ooze, or nannofossil ooze), whereas the siliceous radiolarians, silicoflagellates, and diatoms constitute the siliceous oozes (radiolarian ooze or diatomaceous ooze).

The hemipelagic sediments are basically made of the same components as the deep-sea red clay and the biogenic oozes, clay minerals and biogenic particles respectively, but they also contain an additional and sometimes dominating amount of terrigenous material, such as quartz, feldspars, detrital clay minerals, and some reworked biogenous components from the shelves (Table 1.4, Fig. 1.9).

Most deep-sea sediments can be described according to their composition or origin as a three-component system consisting of

(i) biogenic carbonate,
(ii) biogenic opal, and
(iii) non-biogenic mineral constituents.

The latter group comprises the components of the deep-sea red clay and the terrigenous siliciclastics. On the basis of experiences made in the Deep Sea Drilling Project, Dean et al. (1985) have

Table 1.4 Classification of deep-sea sediments according to Berger (1974).

I. (Eu-)pelagic deposits (oozes and clays)
< 25 % of fraction > 5µm is of terrigenic, volcanogenic, and/or neritic origin
Median grain size < 5µm (except in authigenic minerals and pelagic 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 (or marl) clay
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 terrigenic, volcanogenic, and/or neritic origin
Median grain size >5 µm (except in 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 ~
B. Terrigenous muds, $CaCO_3$ < 30 %. Quarz, feldspar, mica dominant
Prefixes: quartzose, arkosic, micaceous
C. Volcanogenic muds. $CaCO_3$ < 30 %. Ash, palagonite, etc., dominant

III. Special pelagic and/or hemipelagic deposits
1. Carbonate-sapropelite cycles (Cretaceous).
2. Black (carbonaceous) clay and mud: sapropelites (e.g., Black Sea)
3. Silicified claystones and mudstones: chert (pre-Neogene)
4. Limestone (pre-Neogene)

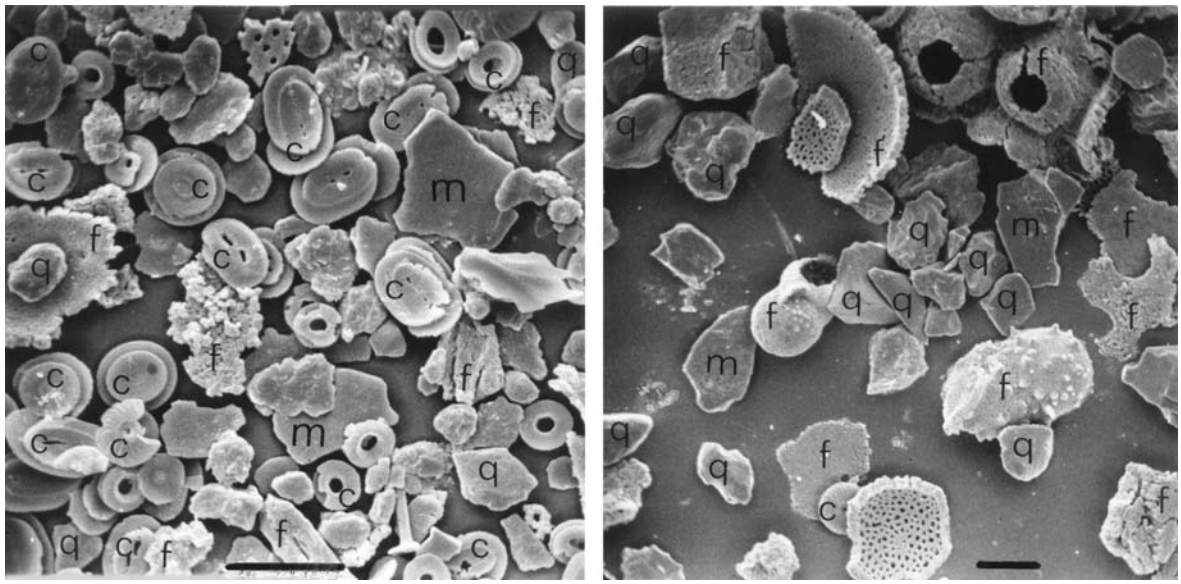


Fig. 1.9 Hemipelagic sediment from Sierra Leone Rise, tropical North Atlantic. *Left*: Fine silt-size fraction composed of coccoliths (c), foraminiferal fragments (f) and of detrital quartz (q) and mica (m). *Right*: Coarse silt-sized fraction predominately composed of foraminiferal fragments (f), some detrital quartz (q) and mica (m), scalebar 10 μm.

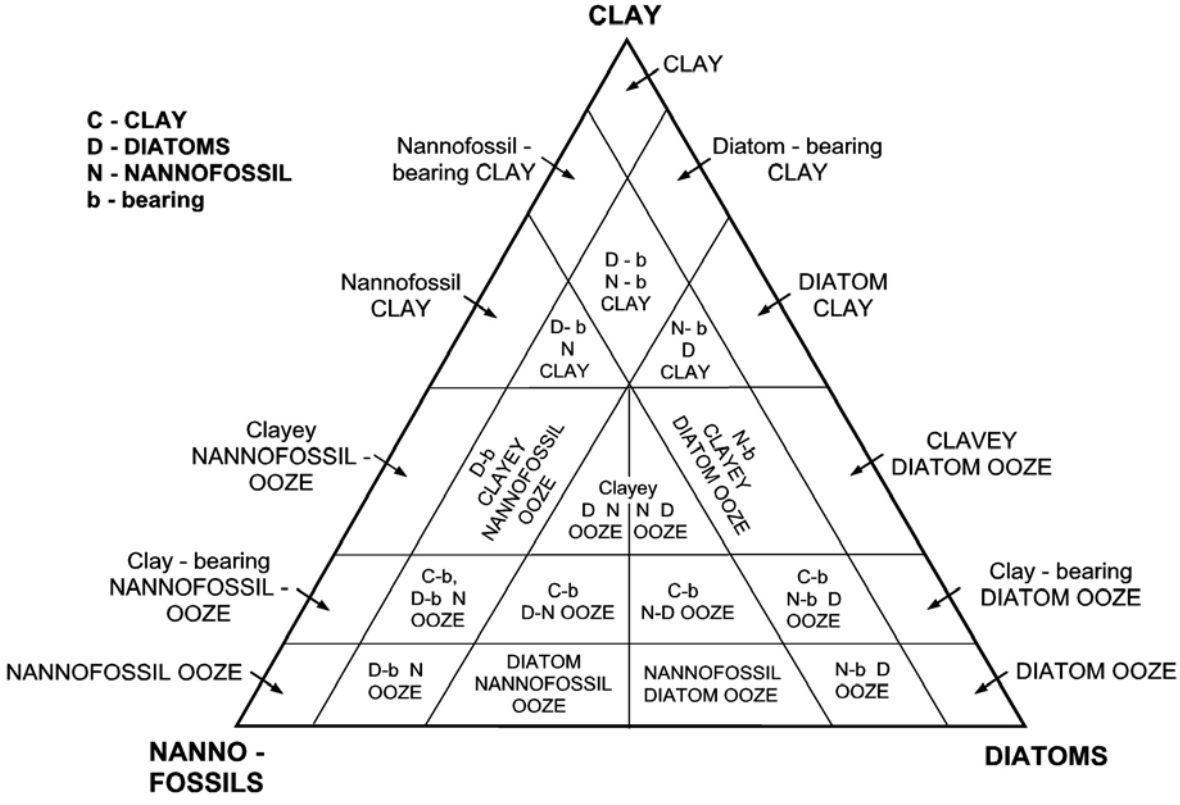


Fig. 1.10 Classification of deep-sea sediments according to the main constituents, e.g. clay (non-biogenic), diatoms (siliceous biogenic), and nannofossils (calcareous biogenic); (modified from Dean et al. 1985).

developed a very detailed and purely descriptive classification scheme (Fig. 1.10):

- The most frequently occurring component with a percentage higher than 50 % determines the designation of the sediment. Non-biogenic materials are specified accordingly on the basis of the grain-size fractions: sand, silt or clay. Biogenic material is referred to as "ooze" and preceded by the most abundant biogenic component: nannofossil ooze, foraminifera ooze, diatom ooze, and radiolarian ooze respectively.
- Each component measuring between 25-50 % is characterized by the following attributes: sandy, silty, clayey, or nannofossil, foraminiferal, diatomaceous, or radiolarian.
- Components with percentages between 10-25 % are referred to by adding the suffix "-bearing", as in "clay-bearing", "diatom-bearing".
- Components with percentages below 10 % are not expressed at all, but may be included by addition of the suffix "rich", as in "C_{org}-rich".

The thus established, four-divided nomenclature of deep-sea sediments with threshold limits of 10 %, 25 % and 50 % easily permits a quite detailed categorization of the sediment which is adaptable to generally rare, but locally frequent occurrences of components, e.g. zeolite, eventually important for a more complete description.

1.4 Global Patterns of Sediment Distribution

The overall distribution pattern of sediment types in the world's oceans depends on few elementary factors. The most important factor is the relative amount with which one particle species contributes to sediment formation. Particle preservation and eventual dilution with other sediment components will modify the basic pattern. The formation and dispersal of terrigenous constituents derived from weathering processes on the continents, as well as autochthonous oceanic-biogenic constituents, both strongly depend on the prevalent climate conditions, so that, in the oceans, a latitude-dependent and climate-related global pattern of sediment distribution will be the ultimate result.

1.4.1 Distribution Patterns of Shelf Sediments

The particulate terrigenous weathering products mainly transported from the continents by rivers are not homogeneously distributed over the ocean floor, but concentrate preferentially along the continental margins, captured either on the shelf or the continental slope (Fig. 1.11). Massive sediment layers are built where continental inputs are particularly high, and preferentially during glacial periods when sea levels were low.

Approximately 70 % of the continental shelf surface is covered with relict sediment, i.e. sediment deposited during the last glacial period under conditions different from today's, especially at times when the sea level was comparatively low (Emery 1968). It has to be assumed that there is a kind of textural equilibrium between these relict sediments and recent conditions. The fine-grained constituents of shelf sediments were eluted during the rise of the sea level in the Holocene and thereafter deposited, over the edge of the shelf onto the upper part of the continental slope, so that extended modern shelf surface areas became covered with sandy relict sediment (Milliman et al. 1972; Milliman and Summerhays 1975).

According to Emery (1968), the sediment distribution on recent shelves displays a plain and distinctly zonal pattern (Fig. 1.12):

Biogenic sediments with coarse-grained calcareous sediments predominate at lower latitudes,

Detrital sediments with riverine terrigenous siliciclastic material at moderate latitudes, and

Glacial sediments of terrigenous origin transported by ice are limited to high latitudes.

In detail, this well pronounced pattern may become strongly modified by local superimpositions. Coarse-grained biogenic carbonate sediments will be found at moderate and high latitudes as well, at places where the riverine terrigenous inputs are very low (Nelson 1988).

Today, as a result of the post-glacial high sea levels, most river-transported fine-grained material is deposited in the estuaries and on the flat inner shelves in the immediate proximity of river mouths. Only a small proportion is transported over the edge of the shelf onto the continental slope. These processes account for the development of mud belts on the shelf, of which 5 types

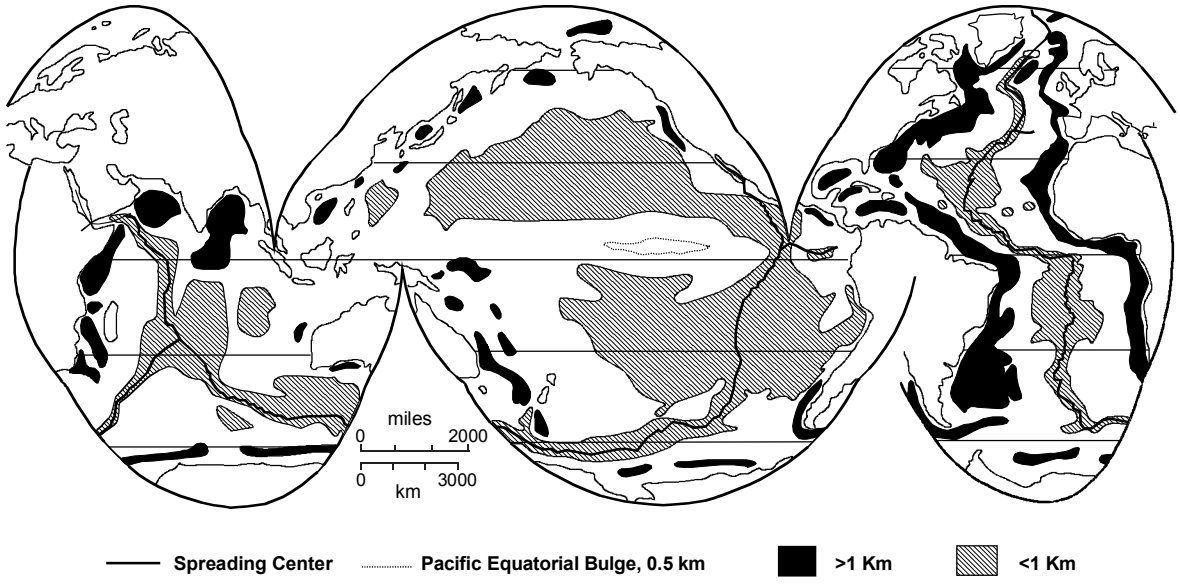


Fig. 1.11 Sediment thickness to acoustic basement in the world ocean (from Berger 1974).

of shelf mud accumulation can be distinguished (McCave 1972, 1985): "*muddy coasts, nearshore, mid- and outer-shelf mud belts and mud blankets*' (Fig. 1.13). Mud belts depend on the amount of discharged mud load, the prevalent tidal and/or current system, or the distribution of the suspended sediment. Muddy coasts will preferentially form near river mouths, whereas mid-shelf mud belts are characteristic of regions where wave and tidal activity are relatively lower than on the inner or outer shelf.

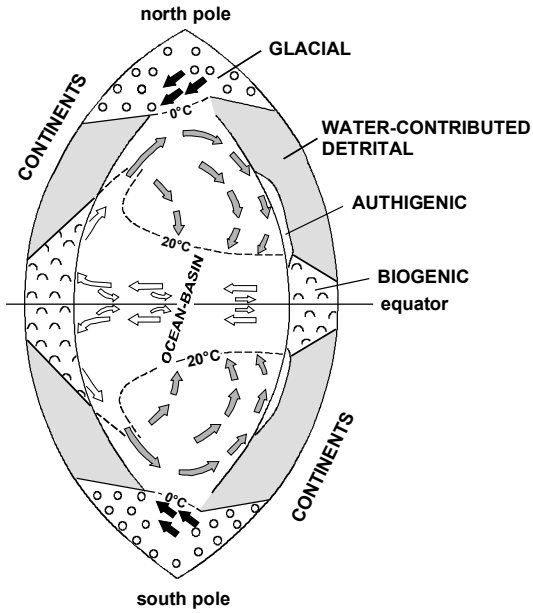


Fig. 1.12 Latitudinal distribution of sedimentary facies of the shallow marine environment of continental shelves in an idealized ocean. *Bold arrows* = cold water; *light arrows* = warm water; *grey arrows* = upwelling water (modified from Reineck and Singh 1973).

Especially in delta regions where the supply rates of terrigenous material are high – as in the tropics – even the entire shelf might become covered with a consistent blanket of mud, although the shelf represents a region of high energy conversion.

1.4.2 Distribution Patterns of Deep-sea Sediments

The two most essential boundary conditions in pelagic sedimentation are the nutrient content in the surface water which controls biogenic productivity and by this biogenic particle production, and the position of the calcite compensation depth (CCD) controlling the preservation of carbonate. The CCD, below which no calcite is found, describes a level at which the dissolution of biogenic carbonate is compensated for by its supply rate. The depth of the CCD is generally somewhere between 4 and 5 km below the surface, however, it varies rather strongly within the three great oceans due to differences in the water mass and the rates of carbonate production.

Calcareous ooze and pelagic clays are the predominant deep-sea sediments in offshore regions

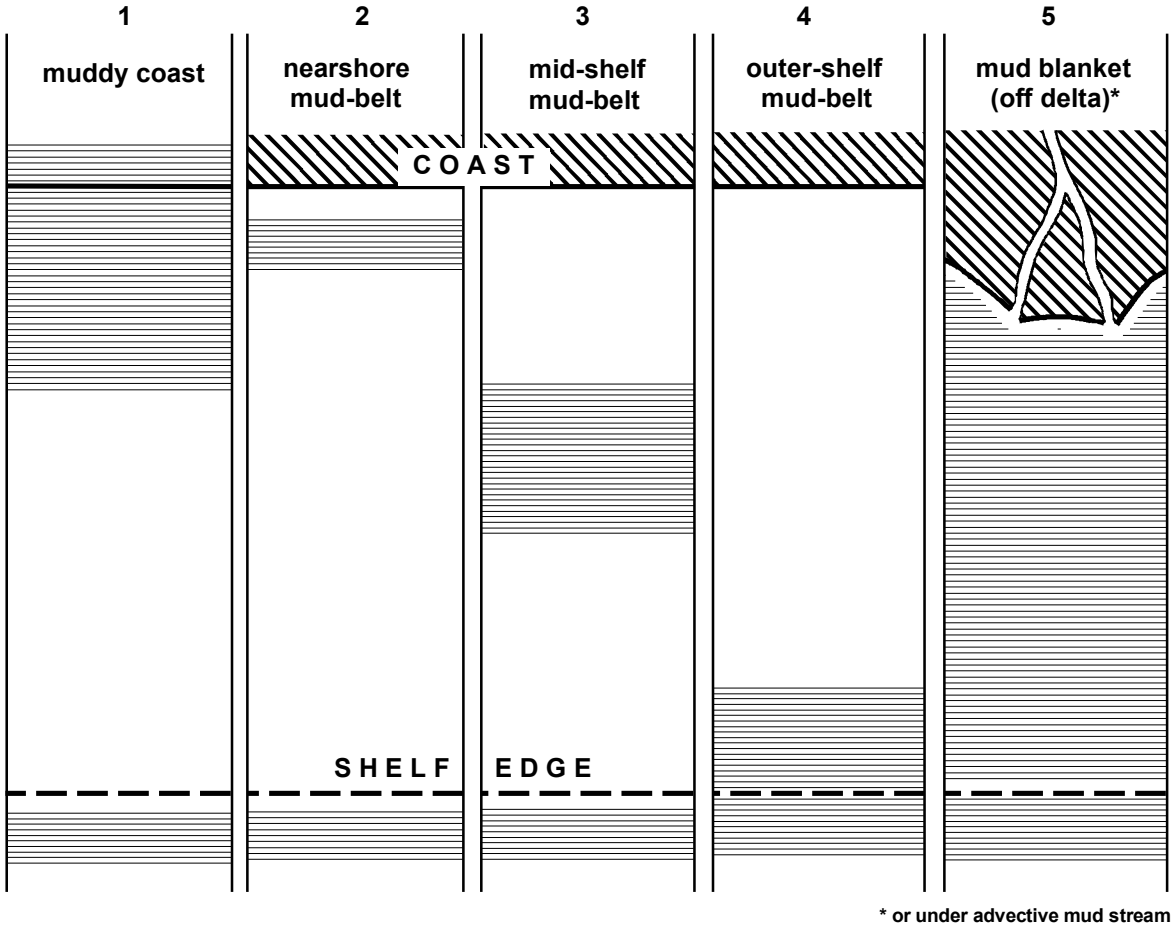
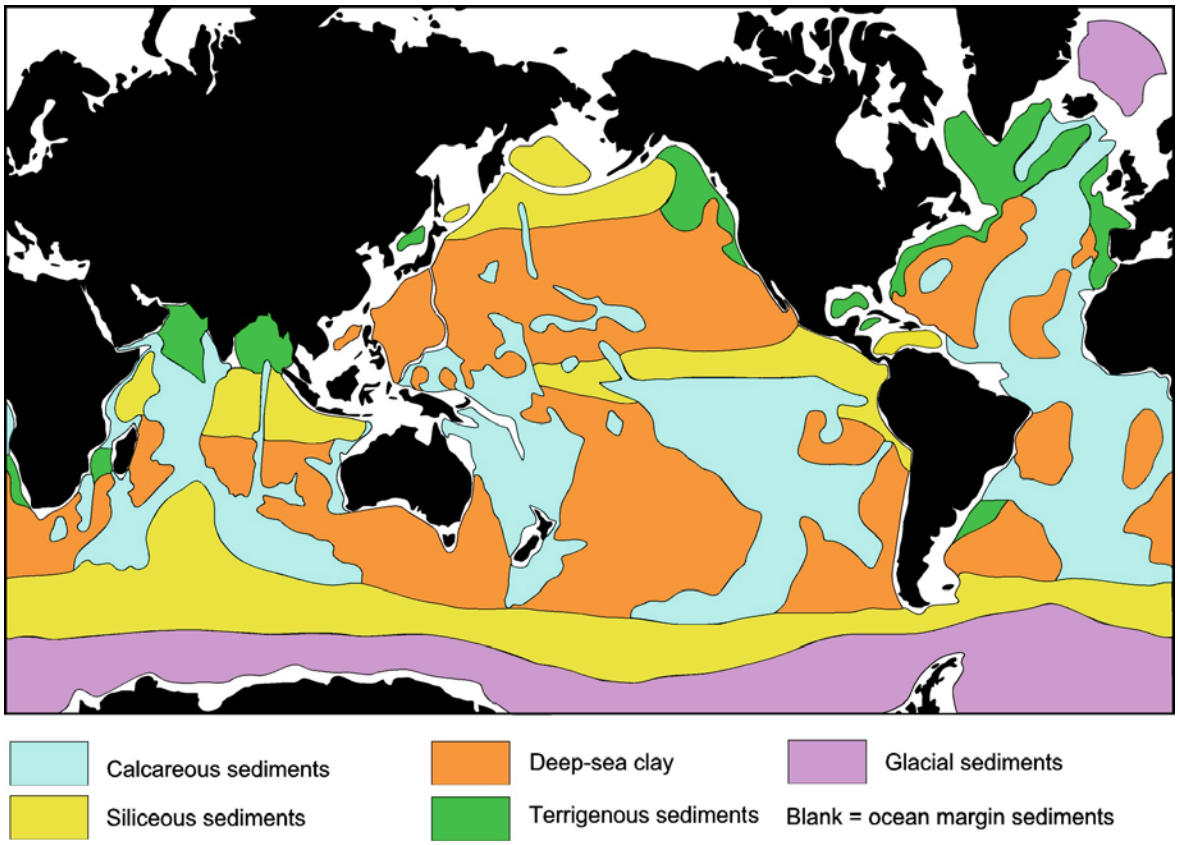


Fig. 1.13 Schematic representation of modern mud accumulation on continental shelves (modified from McCave (1972)).

(Table 1.5). The distribution of these sediments in the three great oceans shows a considerable degree of variation (Fig. 1.14). The distribution patterns strongly depend on the water depth, i.e. the position of the CCD. Calcareous ooze, primarily consisting of foraminiferal oozes and nannoplankton oozes, covers vast stretches of the seafloor at water depths less than 3-4 km and roughly retraces the contours of the mid-oceanic ridges as well as other plateaus and islands, whereas pelagic clay covers the vast deep-sea plains in the form of deep-sea red clay. This particular pattern is especially obvious in the Atlantic Ocean.

Table 1.5 Relative areas of world oceans covered with pelagic sediments; area of deep-sea floor = $268.1 \cdot 10^6$ km^2 (from Berger (1976)).

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
Red clays	25,8	49,1	25,3	38,1
Relative size of ocean (%)	23,0	53,4	23,6	100,0



Fig, 1.14 Distribution of dominant sediment types on the present-day deep-sea floor. The main sediment types are deep-sea clay and calcareous oozes which patterns are predominately depth-controlled. (from Davies and Gorsline (1976)).

Siliceous oozes, mostly consisting of diatom ooze, form a conspicuous ring around Antarctica which clearly marks the zone of the Polar Front in the Antarctic Circumpolar Current. A broad band of radiolarian ooze covers the Pacific Ocean below the equatorial upwelling zone, whilst diatom ooze covers the oceanic margins of the Northern Pacific. Terrigenous sediments, especially in the form of mass flow deposits and turbidites cover vast stretches of the near-continental zones in the North Atlantic, the northeastern Pacific, and the broad deep-sea fans off the big river mouths in the northern Indian Ocean. Glacio-marine sediments are restricted to the continental margins of Antarctica and to the high latitudes of the North Atlantic.

1.4.3 Distribution Patterns of Clay Minerals

Apart from the pelagic clays of the abyssal plains and the terrigenous sediments deposited along the continental margins, primarily or exclusively consisting of clay minerals, various amounts of lithogenous clay minerals can be found in all types of ocean sediments (Table 1.6). The relative proportion of the various clay minerals in the sediments is a function of their original source, their mode of transport into the area of deposition – either by eolian or volcanic transport, or by means of water and ice – and finally the route of transportation (Petschick et al. 1996). In a global survey, it is easy to identify the particular interactions which climate, weathering on the continents, wind patterns, riverine transport, and oceanic currents have with regard to the relative distribution of the relevant groups of clay minerals (kaolinite, illite, smectite, and chlorite).

The distribution of kaolinite in marine sediments (Fig. 1.15) depends on the intensity of chemical weathering at the site of the rock's origin and the essential patterns of eolian and fluvial transport. Due to its concentration at equatorial and tropical latitudes, kaolinite is usually referred

Table 1.6 Average relative concentrations [wt.%] of the principal clay mineral groups in the < 2 □m carbonate-free fraction in sediments from the major ocean basins (data from Windom 1976).

	Kaolinite	Illite	Smectite	Chlorite
North Atlantic	20	56	16	10
Gulf of Mexico	12	25	45	18
Caribbian Sea	24	36	27	11
South Atlanic	17	47	26	11
North Pacific	8	40	35	18
South Pacific	8	26	53	13
Indian Ocean	16	30	47	10
Bay of Bengal	12	29	45	14
Arabian Sea	9	46	28	18

to as the "clay mineral of low latitudes" (Griffin et al. 1968).

Illite is the most frequent clay mineral to be found in ocean sediments (Fig. 1.16). It demonstrates a distinctly higher concentration in sediments at mid-latitudes of the northern oceans which are surrounded by great land masses. This follows particularly from its terrigenous origin and becomes evident when the Northern Pacific is compared with the Southern Pacific. The illite concentration impressively reflects the percentage and distribution of particles which were introduced into marine sediments by fluvial transport. The predominance of illites in the sediments of the Pacific and Atlantic Oceans at moderate latitudes, below the trajectories of the jet-stream, indicates the great importance of the wind system in the transport of fine-dispersed particulate matter.

The distribution pattern of smectite differs greatly in the three oceans (Fig. 1.17), and along with some other factors may be explained as an effect induced by dilution. Smectite is generally considered as an indicator of a "volcanic regime" (Griffin et al. 1968). Thus, high smectite concentra-

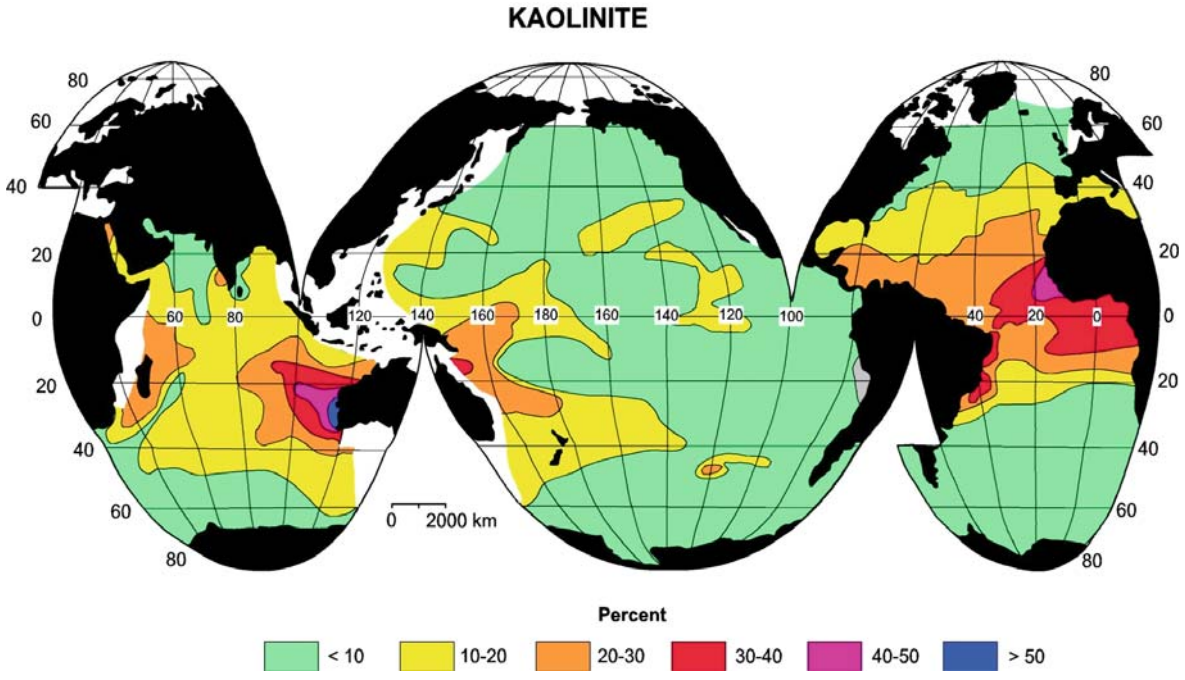


Fig. 1.15 Relative distribution of kaolinite in the world ocean, concentration in the carbonate-free < 2 µm size fraction (from Windom 1976).

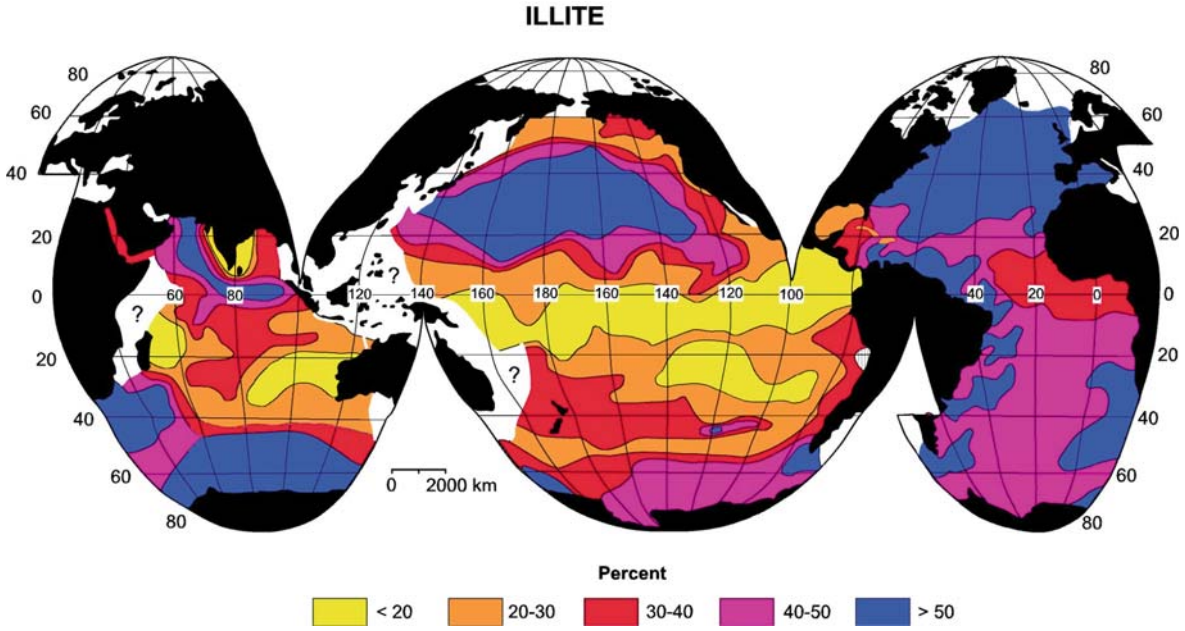


Fig. 1.16 Relative distribution of illite in the world ocean, concentration in the carbonate-free < 2 μm size fraction (from Windom 1976).

tions are usually observed in sediments of the Southern Pacific, in regions of high volcanic activity, where the sedimentation rates are very low due to great distances from the shoreline, and where the dilution with other clay minerals is low as well. The low smectite concentration in the North Atlantic results from terrigenous detritus inputs which are rich in illites and chlorites.

The distribution of chlorite in deep-sea sediments (Fig. 1.18) is essentially inversely related to

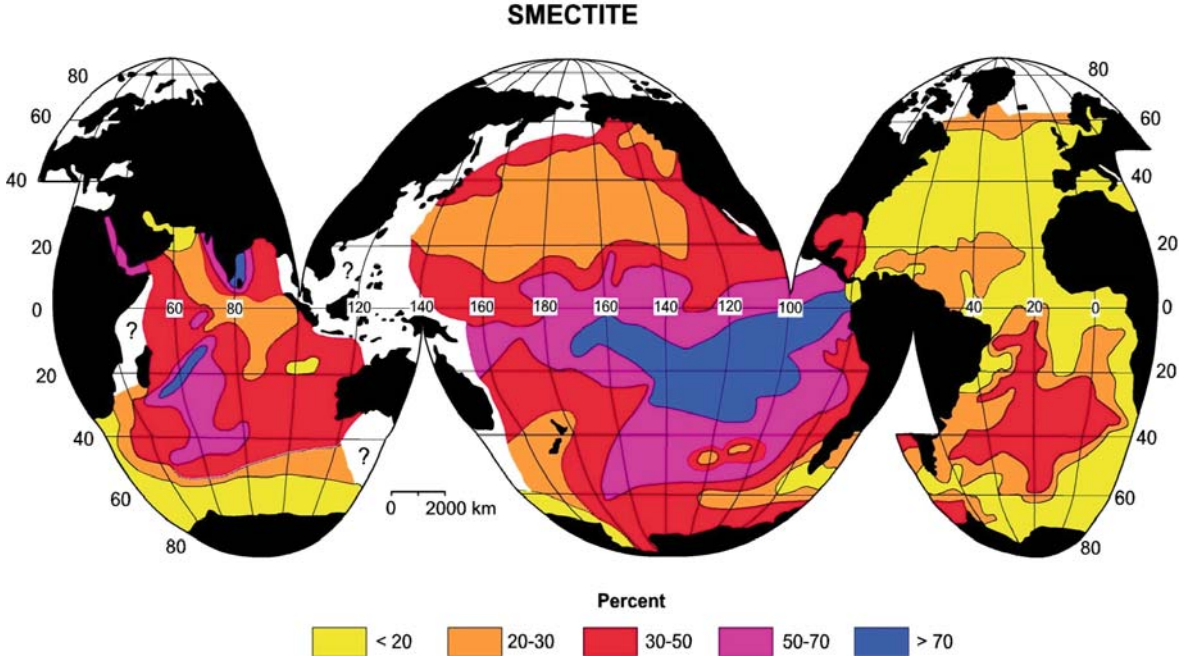


Fig. 1.17 Relative distribution of smectite in the world ocean, concentration in the carbonate-free < 2 μm size fraction (from Windom 1976).

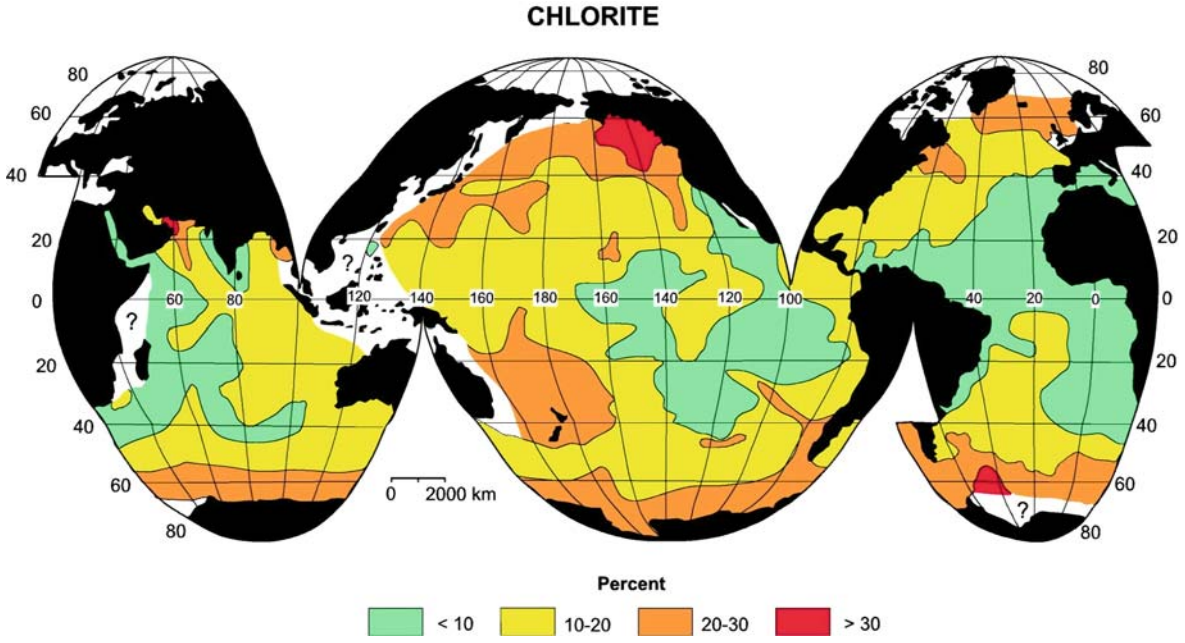


Fig. 1.18 Relative distribution of chlorite in the world ocean, concentration in the carbonate-free < 2 µm size fraction (from Windom 1976).

the pattern of kaolinite. Although chlorite is distributed homogeneously over the oceans, its highest concentration is measured in polar regions and therefore is referred to as the "high latitude mineral" (Griffin et al. 1968).

1.4.4 Sedimentation Rates

As can be seen in Table 1.8, the sedimentation rates of typical types of deep-sea sediments show a strong geographical variability which is based on the regionally unsteady import of terrigenous material and a highly variable biogenic productivity in the ocean.

Table 1.7 Sedimentation rates of red clay in various deep-sea basins of the world ocean (data from various sources, e.g. Berger 1974, Gross 1987).

	Rate (mm ky^{-1})	
	Mean	Range
North Atlantic	1,8	0.5-6.2
South Atlantic	1,9	0.2-7.5
North Pacific	1,5	0.4-6.0
South Pacific	0,45	0.3-0.6

Basically, it can be stated that the sedimentation rate decreases with increasing distance from a sediment source, may this either be a continent or an area of high biogenic productivity. The highest rates of terrigenous mud formation are recorded on the shelf off river mouth's and on the continental slope, where sedimentation rates can amount up to several meters per one thousand years. Distinctly lower values are observed at detritus-starved continental margins, for example of Antarctica. The lowest sedimentation rates ever recorded lie between 1 and 3 mm ky^{-1}. and are connected to deep-sea red clay in the offshore deep-sea basins (Table 1.7), especially in the central Pacific Ocean.

Calcareous biogenic oozes demonstrate intermediate rates which frequently lie between 10 and 40 mm ky^{-1}. Their distribution pattern depends on the biogenic production and on the water depth, or the depth of the CCD. As yet, rate values between 2 and 10 mm ky^{-1}. were considered as normal for the sedimentation of siliceous oozes. Recent investigations in the region of the Antarctic Circumpolar Current have revealed that a very high biogenic production, in connection with lateral advection and "sediment focusing", can even give rise to sedimentation rates of more than 750 mm per thousand years (Fig. 1.19).

Table 1.8 Typical sedimentation rates of recent and subrecent marine sediments (data from various sources, e.g. Berger 1974, Gross 1987).

Facies	Area	Average Sedimentation Rate (mm ky^{-1})
Terrigeneous mud	California Borderland	50 - 2,000
	Ceara Abyssal Plain	200
	Antarctic Continental Margin	30-65
Calcareous ooze	North Atlantic (40-50 °N)	35-60
	North Atlantic (5-20 °N)	40-14
	Equatorial Atlantic	20-40
	Caribbean	~28
	Equatorial Pacific	5-18
	Eastern Equatorial Pacific	~30
	East Pacific Rise (0-20 °N)	20-40
	East Pacific Rise (~30 °N)	3-10
	East Pacific Rise (40-50 °N)	10-60
Siliceous ooze	Equatorial Pacific	2-5
	Antarctic, Indian Sector	2-10
	Antarctic, Atlantic Sector	25-750
Red clay	Northern North Pacific (muddy)	10-15
	Central North Pacific	1-2
	Tropical North Pacific	0-1
	South Pacific	0.3-0.6
	Antarctic, Atlantic Sector	1-2

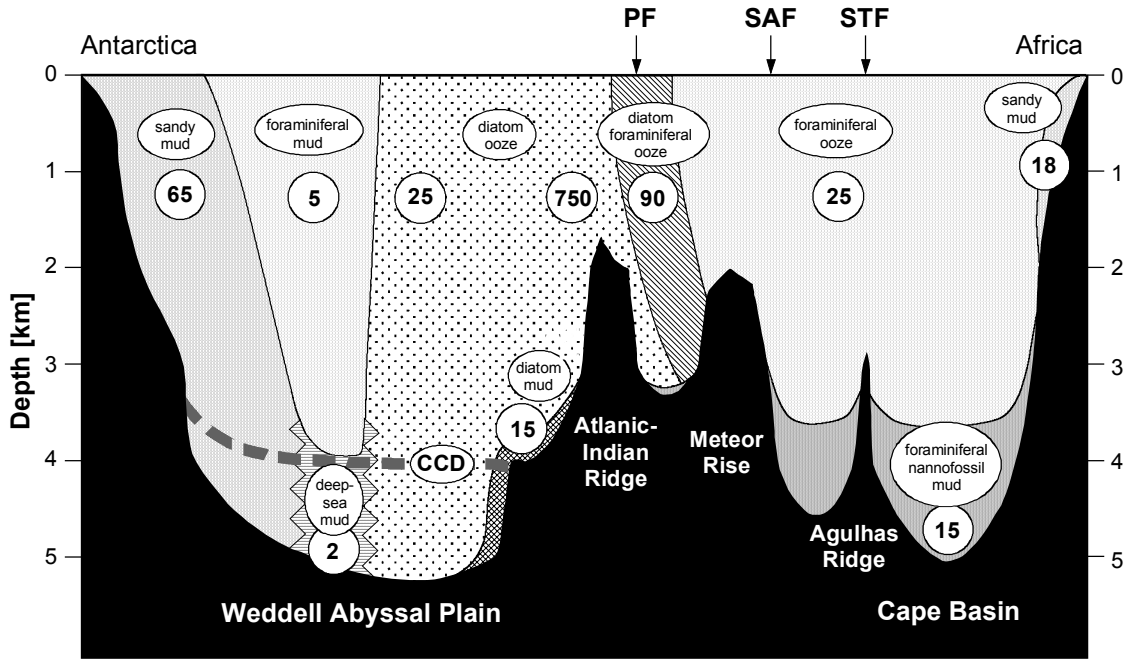


Fig. 1.19 Distribution of major sediment facies across the frontal system of the Antarctic Circumpolar Current (ACC) between Africa and Antarctica. Numbers are typical sedimentation rates in mm ky^{-1}. PF = Polar Front, SAF = Sub Antarctic Front, STF = Subtropical Front.

Velocity of sedimentation or deposition of material through time, i.e. sedimentation rate in this chapter is given solely as millimeter per 1000 years (mm/1000 y or mm ky^{-1}). This describes the vertical thickness of sediment deposited in a certain period of time. It assumes more or less linear and constant processes – not regarding any inconsistency hiatuses – between two datum levels and is, therefore, often referred to as linear sedimentation rate. This term has to be distinguished from mass accumulation rate which refers to a sediment mass deposited in a certain period of time within a defined area, usually given as grams per centimeter square per 1000 years (g cm^{-2} ky^{-1}). For further reading see Bruns and Hass (1999).

1.5 Problems

Problem 1

Is there any correlation between the distribution of calcareous and siliceous sediments, terrigenous sediments and deep-sea clay?

Problem 2

Explain the distribution pattern of clay minerals kaolinite, illite, chlorite and smectite in the world ocean. Why should be illite more common in the North Atlantic than in the South Atlantic?

Problem 3

What are the reasons why sediments of the Atlantic Ocean are thicker along the continental margin than near the mid-Atlantic ridge?

Problem 4

What causes the variability of sedimentation rates?

Problem 5

Summarize and explain the main aspects of deep sea sediment classification.

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